NATURAL ENVIRONMENT RESEARCH COUNCIL

INSTITUTE OF GEOLOGICAL SCIENCES

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REPORT No. 70/15

ICSU/SCOR Working Party 31 Symposium, Cambridge 1970

The Geology of the East Atlantic Continental Margin

3 Europe (continued)

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The Geology of the East Atlantic Continental Margin

3 EUROPE (continued)

Edited by F. M. Delany, Dipl. Geol., F.G.S.

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PREFACE

During March this year a symposium entitled 'Geology of the East Atlantic Continental Margin' was organised by Working Group 31 of the Scientific Committee on Oceanic Research, itself a committee of the International Council of Scientific Unions. The principal aims of the Symposium were

1. to review present knowledge and plans for investigation of the East Atlantic Continental Margin, and
2. to identify gaps in knowledge that could be filled by co-ordinated national or international marine research programmes.

Ten invited general lectures outlined continental margin problems. These were given by specialists in geology, geophysics, oceanography and biology, as well as economic aspects. Nineteen invited regional lectures gave reviews on topography, sediments, rocks and structures of the entire East Atlantic continental margin. Dr. G. Giermann of Intergovernmental Oceanographic Commission/UNESCO summarised planned national and international programmes. Twenty shorter regional contributions of general interest were included.

All but six of these papers are now published in English in four numbers of the Report series: Report 70/13 contains general and economic papers, Reports 70/14 and 70/15 papers concerning the continental margin of Europe and 70/16 papers on the East Atlantic continental margin of Africa. These papers were edited for publication by Miss Frances Delany.

The Institute of Geological Sciences, of course, accepts no responsibility for the views expressed in these papers, or for the diagrams here presented, except where their authors are members of its own staff.

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Director

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October, 1970.

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Vertical crustal movements in the North Sea area through geological time

B. J. Collette

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Vertical crustal movements in the North Sea area through geological time

B. J. COLLETTE

INTRODUCTION

The structure of the North Sea basin has mainly been determined by vertical movements. Since Permian times a subsidence of 6000-7000 m occurred in the deeper parts. The folding that can be observed hardly deserves that name; undulating is a more appropriate adjective for the structures that are encountered.

Although the North Sea is the site of intensive structural research, the explorational character of the investigations prohibits full publication of the results. Apart from publications on the geology of the adjoining countries, we therefore have to rely on several general papers that have been released (Bartenstein, 1968; Hinz, 1968; Kent, 1967 and 1968; Heybroek and others, 1967; Hornabrook, 1967). Recently a progress report was published by Kent and Walmsley (1970), releasing several well data. A really new contribution to the understanding of the structure of the North Sea was made by O. Holtedahl (1970) regarding the Norwegian Channel and, generally, the problem of the 'marginal channels'. The present evidence favours the interpretation that the channels are erosional features (compare Grant, 1966) along the flexure or fault that separates the older core of the continent or subcontinent from the younger strata.

Starting from a regional gravity survey (Collette, 1960) and incorporating the results of a series of seismological experiments in the North Sea (Collette and others, 1967, 1970) and the available structural data, we proposed a few years ago a model accounting for the subsidence of the North Sea basin by a process of crustal thinning by phase transformation (Collette, 1968). Instead of summarising all published structural data and repeating previous papers, we refer to the literature mentioned. In this paper we shall consider the North Sea area from one aspect only, the problem of the vertical movements. Subsidence and upheaval are generally accepted ideas in geology in terms of description. With the exception of several outstanding phenomena, like the postglacial upheaval of Fennoscandia and a single clear-cut case of graben formation, their cause is not well understood, however. And, again with several exceptions, as a geophysical problem subsidence and upheaval not related to mountain building processes (folding belts) do not receive the attention they deserve. This may partly be caused by the lack of models into which the structural information is condensed. We shall try to give such a model for the North Sea area and indicate which of the recorded movements can be explained by known processes or acceptable hypotheses, and which cannot.

THE STRUCTURAL SETTING

The North Sea basin is surrounded by several older structural units: the Caledonian mountain system of Scotland and Norway to the north-west and north, the Precambrian of South Norway to the north-east and the Hercynian mountain system to the south (Fig. 1). We believe that the Caledonian system of Norway and Scotland are not directly connected. The structural situation in the northern part of the North Sea would be comparable to that in the Barents Sea to the north: in both cases the Caledonian system bifurcates. The Shetlands form a clear indication for north-south structures in the subsurface. The missing part of the reconstructed Caledonian system can be found in East Greenland (Harland, 1958; Kay, 1967), which drifted away after the opening up of the Atlantic. Speaking in terms of continental drift and accepting this concept also for the past, the Caledonian folding resulted from a collision of continents, comparable to the Tethys-zone of the Alpine system. The situation in the late Palaeozoic is less clear. However, we would venture the suggestion that the Hercynian system in Western Europe was comparable to the circum-Pacific part of the Alpine system and actually formed a cordillera, bounded to the south by an ocean.

The North Sea basin and the affiliated north-west German basin developed in the space left free by the Caledonian and Hercynian orogenies, c.q. affected by these orogenies to a lesser degree. We do not know what the underground really consists of. Older orogenies may have left their traces. Nevertheless, it does not seem preposterous to suppose that a relation exists: the Caledonian and Hercynian orogenies will have had their impact on the chemical composition of the lower part of the earth crust; the changed chemical composition might be less suited for phase transitions of the type we believe to occur under the North Sea basin.

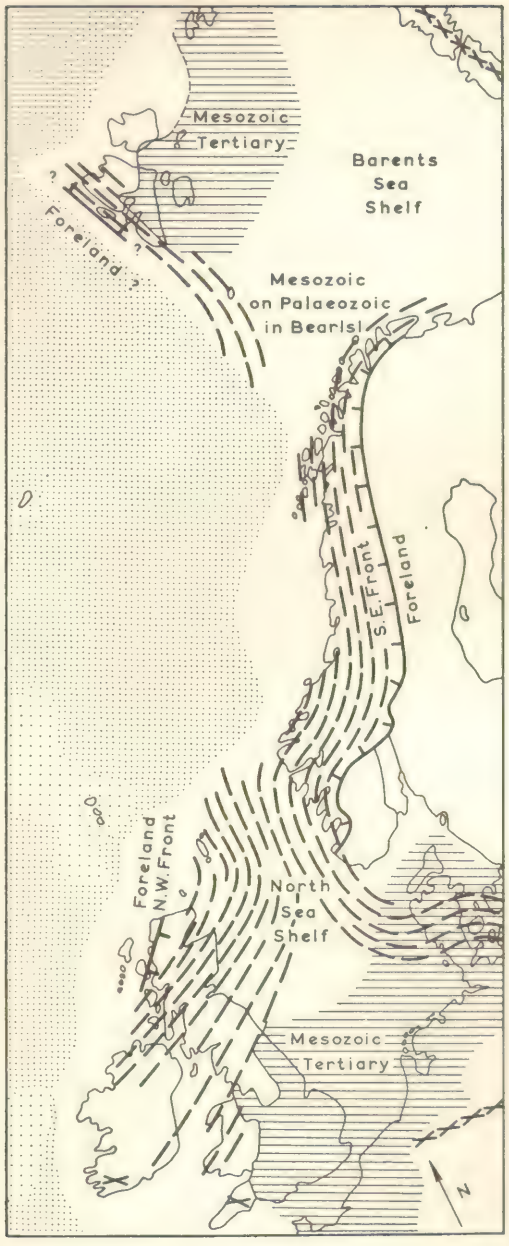

Fig. 1. Structural setting of the North Sea basin

A far more important point, however, is the following. One frequently comes across the opinion that the uplands of Scotland and Norway and the Ardennes in Belgium represent the remnants of the Caledonian and Hercynian orogenies. This cannot be true, since denudation is too fast a process to allow a mountain system to live that long. Actually, there have been several periods of peneplanation since, the probably best known occurring during the Early Permian (the Rotliegendes). The formation of the north-west European uplands is of much younger date (late Caenozoic?), the Ardennes as an uplifted peneplain dating from the late Tertiary. O. Holtedahl (for example, Holtedahl and others, 1960) drew attention to the high elevation of the northern coasts in general: north-west Europe, Greenland and north-east America. The crustal thickening involved is enormous: for the Scandinavian mountains it is of the order of magnitude of 10 km (2 km plus 4.5 times 2 km isostatic compensation). The question of the origin of this huge volume of light crustal rock is still unanswered. Accepting continental drift as a most important process in shaping the Earth's surface, the question can also be formulated as follows: why do we find both erosional and depositional coasts along the Atlantic (McGill, 1958)?

In terms of vertical movements, the upheaval of the northern coasts is as important as the subsidence of the North Sea basin. Trying to answer it with our present knowledge would be pure guess-work. Instead a short comment is made on the term 'orogeny'. Its ambiguity became clear when discussing the north-west European uplands; the recent process of upheaval certainly led to the formation of mountains. Nevertheless, it is not a mountain building process in the usual sense. One might consider abolishing the term 'orogeny' and related terms and instead use neutral words as 'folding' and 'upheaval'. This would be in line with Rutten's use of the terms 'uplands' and 'mountain chains' in his recent book, 'The Geology of Western Europe' (1969), to distinguish between uplifted plateaus ('basement') and young folding belts.

THE CARBONIFEROUS IN NORTH-WEST EUROPE

Although the Caledonian and Hercynian folding did not affect the North Sea area, one aspect of its Palaeozoic history should be mentioned: the Carboniferous sedimentation. A much larger area than covered at present by the North Sea basin underwent a subsidence of several kilometres. The evidence of the surrounding countries points to the existence of one large basin. Whether the Carboniferous is actually still present everywhere under the North Sea basin we ignore. The Hercynian folding came to a halt to the south of what is now the Brabant Massif. As an elevated structure the Brabant Massif developed only after the folding. Nevertheless, it may well have acted as a subsurface pressure barrier, the total thickness of Devonian and Carboniferous sediments to the south being greater than to the north (compare Collette, 1960, p. 66). In our concept the North Sea area then formed part of a continental shelf, originally bordered to the south by an ocean. The Hercynian mountains ('Meso-Europe') were shoved against this shelf like the Rocky Mountains against the older core of the American continent.

Here there is no opportunity to elaborate this idea or check its geological implications.

Fig. 2. Depth indications of base Zechstein. After Heybroek and others (1967).

Reproduced from the Proceedings of the 7th World Petroleum Congress, Mexico, 1967. Elsevier, Barking

However, the suggested configuration should be considered seriously when in future a reconstruction is made of the pre-Hercynian position of the continents, that is, one step further back into history than the present reconstructions go. For our problem it is only important to know that the North Sea area has been involved in an earlier subsidence. What happened to the area before this subsidence is not known and cannot be found by extrapolation from the surrounding countries .

THE SUBSIDENCE OF THE NORTH SEA BASIN

When speaking of the North Sea basin, we only consider the post-Hercynian history.

After the deposition of the red beds of the Lower Permian, the Rotliegendes, which generally is interpreted as a peneplanation phase, the development of the Zechstein basin started a history of subsidence (Fig. 2) which is still going on. The subsidence was not continuous in space and time. Locally, sub-basins sunk faster and parts of the basin were temporarily uplifted and underwent erosion. Nevertheless, the result is a vast basin with a depth of 6000 to 7000 m at the deepest parts. This gives an average rate of subsidence of 0.003 cm/year. Locally the rate of subsidence was much higher, 0.05 cm/year and probably even more. The maximum of the Tertiary subsidence (over

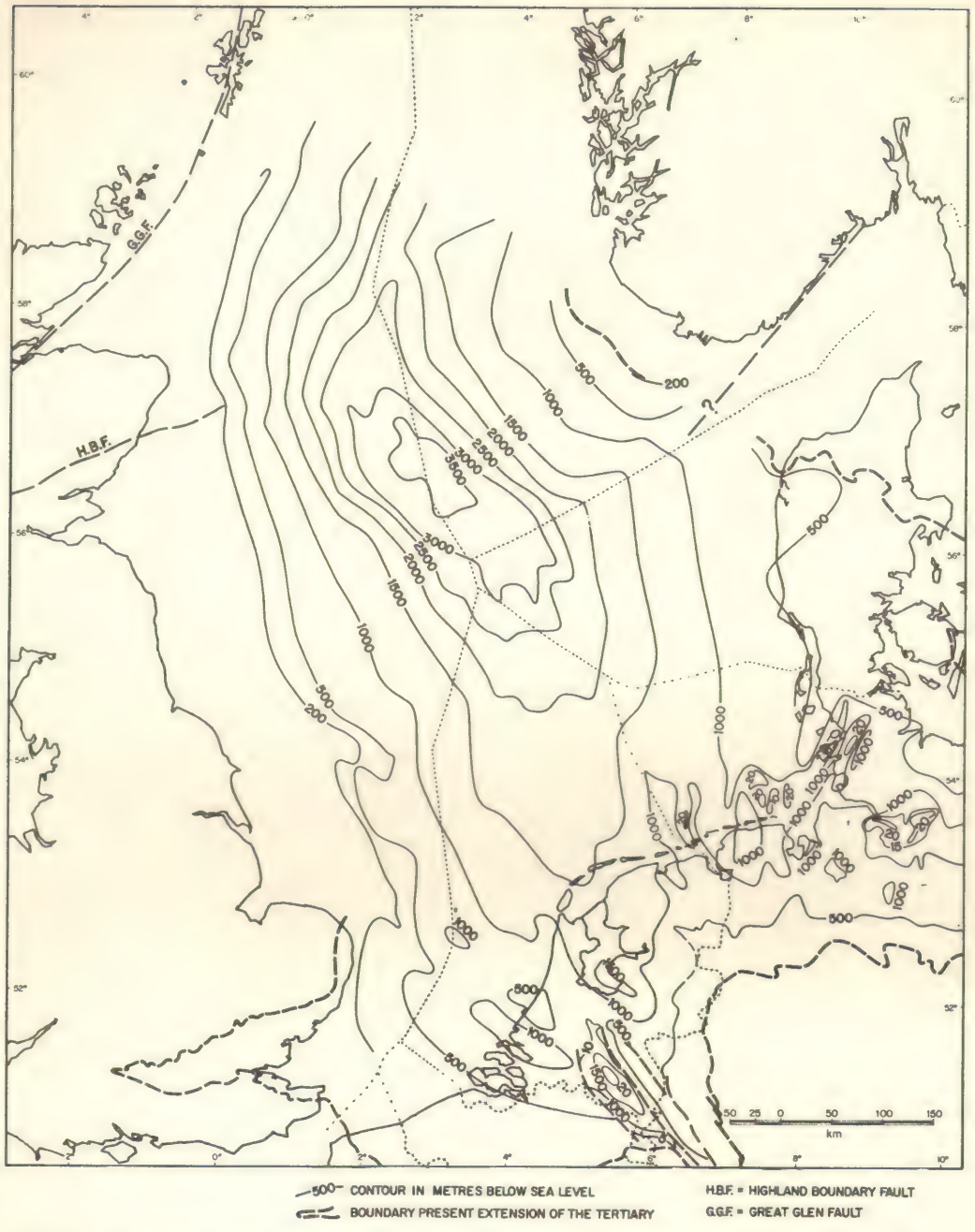

Fig. 3. Contour map of base Tertiary. After Heybroek and others (1967).

Reproduced from the Proceedings of the 7th World Petroleum Congress, Mexico, 1967. Elsevier, Barking

3500 m) is shifted northward with regard to the Zechstein basin (Fig. 3).

Several considerations, which will not be repeated here (see Collette, 1968) led us to believe that during the subsidence the isostatic equilibrium was maintained. This implies that the crust underwent a thinning of roughly 10 km (Fig. 4).

Generally, the sediments are shallow-water deposits, and from this point of view it seems reasonable to interpret the subsidence as a result of a phase change caused by the continuous loading of the crust with sediments. In Collette (1968) I presented a model, starting with an overload of a few hundred metres of sediment (the equivalent of 10 to 20 milligals which can easily be hidden in the 'regional' of the otherwise isostatically compensated area) and involving a phase change in the lower part of the crust. I computed that the heat production of the phase change is small compared to the heat needed for warming up the sinking crustal column and is negligible compared to the crustal heat flow. The process comes to a stop if the crust no longer has the right chemical composition. Parts of the crust converted into heavy, eclogite-type rock (originally we held the basalt-eclogite

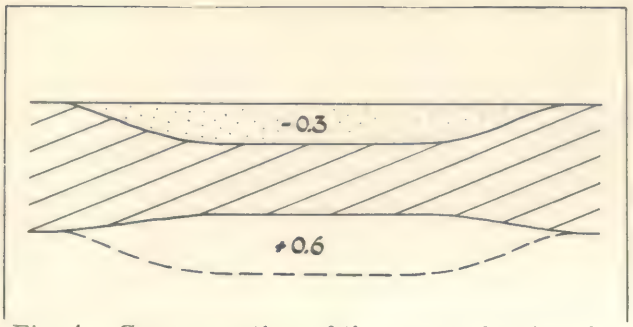

Fig. 4. Cross-section of the crust showing the mass deficiency of the sedimentary column in the North Sea basin and the amount of crustal material that disappeared.

transition responsible for the density change; in view of recent geochemical developments we prefer to be less explicit on the minerals involved) might break free and sink into the less dense mantle. The chemical composition of the mantle would differ from the composition of the lower crust and the process would not necessarily be reversible.

Although we are convinced that non-mechanical crustal thinning is the main process causing the subsidence as plastic thinning would lead to large-scale rifting at the surface, we do not know whether all details can be explained in this way. Some rifting did occur (for example, the Peelgraben on the north flank of the Brabant Massif). Also, some authors (for example, Schmalz, 1969) claim that the Zechstein evaporites have been deposited in medium deep water, considering the high rate of sedimentation ('varves' in rock salt indicate a deposition of 600 m in 12 000 years, or a rate of 5 cm/year). This rate of sedimentation is much higher than the figures mentioned earlier (0.003 to 0.05 cm/year), but since the subsidence would be load-governed, high rates of subsidence are equally well possible. The real question is what depth actually occurred and as far as we can judge, this can be any depth below the wave action, provided that a process is found to account for the high salinity. However, we do not feel qualified to give a definite answer to this problem.

DIFFERENTIAL VERTICAL MOVEMENTS

When a column of sediment sinks, it eventually attains a temperature at which a eutectic granitic melt originates (Nieuwenkamp, 1948). This melt has a lower density than its parent rock and will form an unstable layer (Bott, 1961). The melt will concentrate (gneissic dome), break through the overlying layers (diapiric stage) and spread out again at its equilibrium depth (intrusive stage). The process can be compared to salt tectonics (salt pillow formation, diapirism, mushroom structures), though the dimensions are larger as a result of the greater depth at which the melt originates compared to the depth at which salt becomes unstable. At the surface this will result in the formation of ridges with 'rim geosynclines' (dome and diapiric stages), and collapsing of the ridge combined with (relative) upheaval of the 'rim geosynclines' (intrusive stage).

We proposed this mechanism in our 1968 paper to explain the differential vertical movements that occurred in the North Sea area: Kijkduin High, Smiths Knoll High, Wealden anticline. These are all highs that subsided first and were uplifted later, the net effect being that the base of the basin is now roughly flat. The mid-Netherlands ridge underwent an opposite movement: a maximum thickness of Tertiary sediments is found over this Mesozoic 'ridge'. Voigt (1962) proposed the term 'Randtröge' for these features; the phenomenon would only occur along the Hercynian massifs. We feel that this term is unnecessarily restrictive but refrain from proposing a different term.

With continued subsidence and sedimentation the process might repeat itself. Geochemical experiments should decide whether the remaining rock, that is, the part that does not melt eutectically, thus becomes more suited for phase transitions and which transitions these might be.

CONCLUDING REMARKS

In the foregoing we sketched in broad outlines the development of the North Sea basin against the setting of Caledonian and Hercynian mountain building periods. It was argued that the elevation of the Scottish and Norwegian uplands and of the Ardennes is a relatively young feature, which has no causal relation with the Caledonian and the Hercynian folding. A description is given of a model proposed a few years ago (Collette, 1968) of the subsidence of the North Sea basin, involving crustal thinning by phase transition in the lower part of the crust under the increasing load of sediment. Parts of the crust that became heavy in this way, may break free from the rest of the crust and disappear into the mantle. Differential vertical movements are explained as a result of the forming of an unstable granitic layer in the crust itself.

We are painfully aware that this model solves only half of the problem: it does not account for the upheaval of the uplands. We do not see an easy solution for this aspect. The absence of a geographic relation between the present uplifts and the older orogenies, as also clearly demonstrated in Greenland, rules out the possibility of a long-term after-effect of the involved orogenies on the upper part of the mantle (compare Hsü, 1965).

Heat or light material should be transported to the areas that are uplifted. Mantle convection might be the agent, but we do not have a single proof or even indication that this is actually so. For the moment we therefore prefer to stop the discussion at this point.

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Sediments of the southern North Sea

H. J. Veenstra

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SUMMARY

Holocene sediments of the North Sea floor discussed here are less than 10 m thick in most cases and consist of mud (< 50 μ), sand (50 μ - 2 mm) and gravel (> 2 mm).

The study of heavy minerals by previous workers suggests that the majority of the sand in the North Sea has a glacial origin, but that near the south-western coast of Holland and north of Dover Strait the sand has been brought in by an old course of the Rhine. This is further confirmed by roundness studies of quartz grains.

The grain size of the sands shows a decrease to the north, which could indicate that currents have transported the fines in that direction.

The sand waves in the Southern Bight do not migrate appreciably, but probably are modified in shape by storms.

The fine mud that enters the tidal flats of the North Sea mainland coasts consists chiefly of suspended matter from the River Rhine. It is probable that the mud of the North Sea floor has the same source, although outwash from submarine moraines may have contributed some material.

The gravels of the sea floor can be subdivided into four groups according to grain size, composition and roundness. The first type consists of reworked fluvial gravel deposited by Pleistocene rivers off the south-eastern British and the Belgian coasts. The other three gravel groups are of glacial origin.

SOMMAIRE

Les sédiments au fond de la Mer du Nord sont discutés. Ils sont pour la plupart moins de 10 m et consistent en vase (< 50 μ), sable (50 μ -2 mm) et gravier (>2 mm).

Des études antérieures de minéraux lourds suggèrent que la plupart du sable dans la Mer du Nord provient d'origine glaciaire, mais que, près de la côte du sud-ouest de la Hollande et au nord du Pas de Calais, le sable a été apporté par un cours ancien du Rhin. D'autre part des études sur la rondeur de grains de quartz l'ont confirmé.

Les grains des sables subissent une diminution vers le nord, ce qui implique la possibilité que des courants ont apporté les grains les plus fins en cette direction.

Les vagues de sable dans la Baie du Sud ne migrent pas à un degré appréciable, mais elles sont probablement modifiées en configuration par des tempêtes.

La boue fine qui entre les wadden de la côte continentale de la Mer du Nord consiste sourtout en particules en suspension venant du Rhin. Il est probable que la boue au fond de la Mer du Nord a la même source, mais elles peuvent aussi venir en quelque part des moraines sousmarines.

Les graviers du fond de la mer se subdivisent en quatre groupements, soit par granulométrie, par composition et par roundeur.

Le type premier se compose de graviers fluviatiles remaniés et déposés par des fleuves pleistocènes le long des côtes du sud-est de la Bretagne et de la Belgique. Les autres trois groupements de gravier sont d'origine glaciaire.

ZUSAMMENFASSUNG

Besprochene Holozänische Sedimente vom Nordseeboden haben meistens eine Dicke von weniger als 10 m und bestehen aus Schlamm ($< 50\,\mu$), Sand ($50\,\mu$ -2 mm), und Kies (> 2 mm).

Die wissenschaftliche Arbeit über schwere Minerale von früheren Arbeitern suggeriert, dass der Sand in der Nordsee meistenteils einen Gletscherursprung hat, aber dass der Sand in der Nähe von der südwestlichen Küste von Holland und nördlich von der Dover-Strasse von einem alten Gang des Rheins hereingebracht war.

Die Untersuchung von der Rundheit der Quarzkörne bestätigt weiter diese Meinung.

Die Korngrösse der Sände zeigt eine Verminderung zum Norden, dass bedeuten könnte, dass Strömungen die feinsten Sände in jene Richtung getragen haben.

Die Sandwellen im südlichen Bucht wandern nicht bemerkbar; ihre Bildung ist wahrscheinlich von Stürmen abgeändert.

Der feine Schlamm, der in die Watten vom Nordseefestland eingeht, besteht hauptsächlich aus suspendiertem Stoff vom Rhein. Wahrscheinlich hat der Schlamm vom Nordseeboden den selben Ursprung, obgleich es möglich ist, dass Ablagerung von Unterwassermoränen etwas Stoff beigetragen hat.

Man kann die Kiese des Seebodens in vier Gruppen ihrer Grösse, ihrer Bildung und ihrer Rundheit entsprechend einteilen. Die erste Typ besteht aus aufgearbeiteten Fluvialkiesen fort von den Südostbritischen- und den belgischen Küsten abgelagert. Die anderen drei Kiesgruppen haben einen Gletscherursprung.

Sediments of the southern North Sea

H. J. VEENSTRA

INTRODUCTION

The North Sea forms an old sedimentary basin with thick Mesozoic and Caenozoic beds that were, in most cases, laid down in shallow water. During the Tertiary the basin began to assume roughly its present shape. Due to tectonic movements including the diapiric intrusion of Permian salt the thickness of Caenozoic sediments in the North Sea varies considerably, but generally does not exceed 3500 m (Hinz, 1968). The Tertiary beds may reach a thickness of 2500 m, the Pleistocene beds 1000 m and the Holocene 10 m (Fig. 1).

Tertiary sediments as clay and crags crop out in the extreme southern part of the North Sea. Although Pleistocene sediments crop out in some places in the central and eastern part of the North Sea, the major part of the sea floor consists of Holocene sediments a few metres thick.

A practical subdivision of the sediments found at the North Sea floor would be mud (grain size $< 50\ \mu$), sand (grain size $50\ \mu$ - 2 mm) and gravel (grain size > 2 mm) (Fig. 2).

MUD

Mud includes all matter of which the rate of fall in still water is less than that of a quartz grain $50\ \mu$ in diameter (Terwindt, 1967). It consists of silt and clay and possesses the following components: quartz and feldspar fragments; mica flakes; clay minerals; carbonates; organic components.

The carbonates are partly derived from molluscs, foraminifera or echinoderms. Sources of the other mineral components are rivers, boulder clays and old clay deposits.

Rivers

The yearly discharge of many rivers has been measured during several decades and the quantities of fine suspended material in the rivers are also known. The following list, partly after De Groot (1963) and Terwindt (1967), shows the quantities of dry matter transported to the North Sea:

	tons/year
Strait of Dover	2×10^6
Rhine	4×10^6
Meuse	1×10^6
Scheldt	0.1×10^6
Thames	0.2×10^6
Humber	0.1×10^6
Ems	0.1×10^6
Weser	0.6×10^6
Elbe	0.7×10^6
Total	9×10^6

De Groot (1963) studied the manganese compounds of the muds deposited in the tidal flats along the Dutch and German coasts. It appeared that suspended material from the Rhine that was brought into the tidal flats formed the dominant part of the fine fraction. Terwindt (1967) confirms this view stating that the zone of mud transport close to the Dutch shore north of Hook of Holland is about 2 km wide.

Boulder clays

It appears that morainic ridges occur on the North Sea floor (Pratje, 1951; Veenstra, 1965; Oele, 1969). A rough estimate shows that 60 000 km^2 of the North Sea floor is covered with gravel, probably belonging to eroded boulder clay deposits.

It is certain that boulder clay of Saalian age as well as of Weichselian age crops out on the North Sea floor (Veenstra, 1969). In a 100 km wide belt in front of the British and Danish coasts patches of Weichselian boulder clay are found, whereas off the Frisian Islands Saalian boulder clay occurs. The boulder clay cliffs of East Anglia consist chiefly of Saalian boulder clay, but in Yorkshire Weichselian boulder clay predominates in the cliffs.

The sand and clay fractions that have been winnowed out by waves and currents are redeposited in the North Sea. The average content of sizes $< 40\ \mu$ amounts to 30 per cent in boulder clays of north-west Europe. The boulder clays may have been an important source of mud in the beginning of the Holocene, when they were subject to intense wave action and not covered by younger sediments.

Old clay deposits

Old clay deposits that could have supplied mud outcrop in some places on the sea floor, in the tidal inlets between the Frisian Islands and at eroded cliffs.

In the southern part of the North Sea, Eocene clay may reach the sea floor, whereas near the Dutch and German coasts Atlantic clay may crop out.

In the tidal inlets Eemian clay crops out, whereas in eroded cliffs of ice-pushed sediments in Schleswig-Holstein and Jutland interglacial

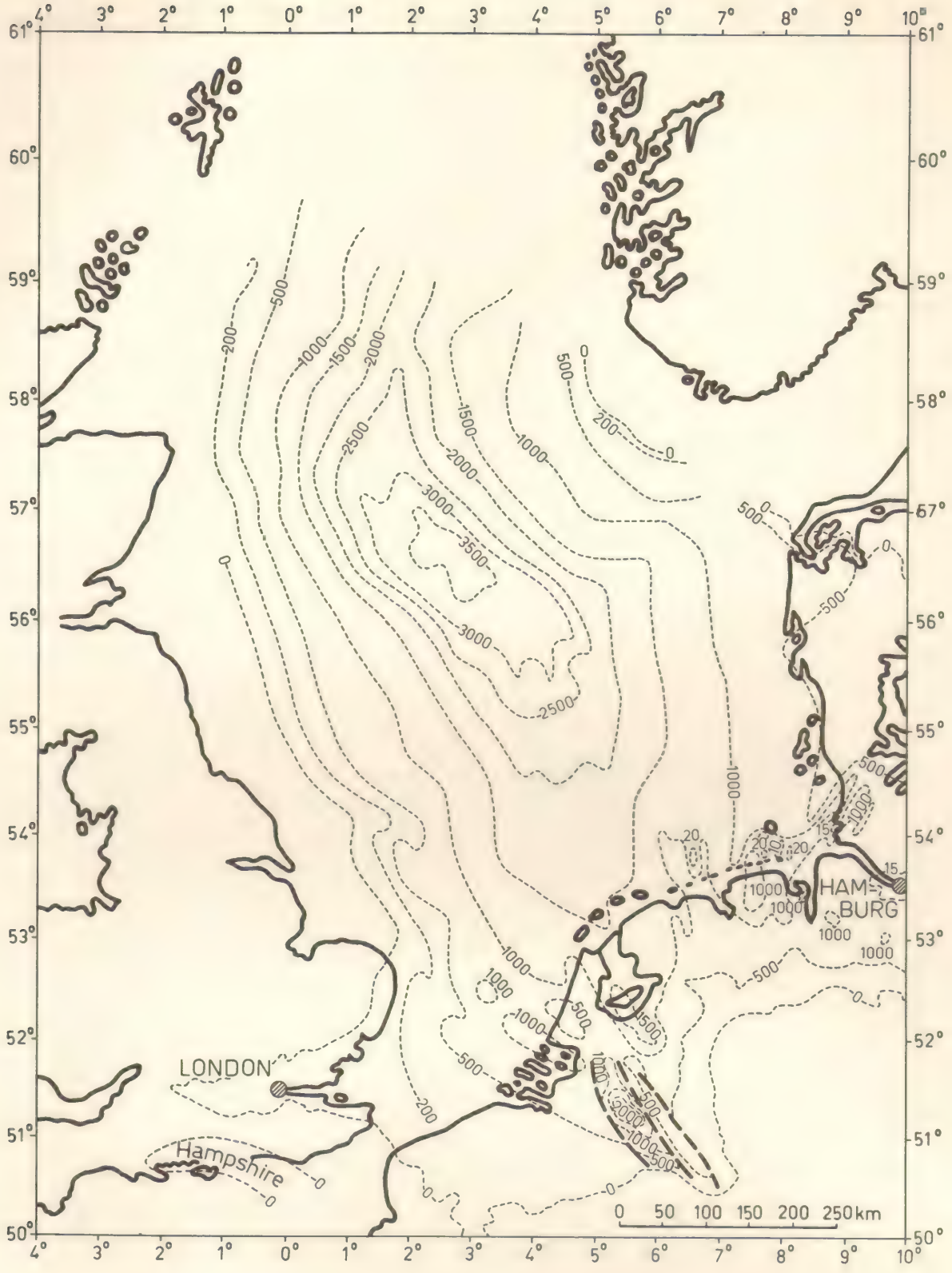


Fig. 1. Thickness of Caenozoic sediments in the North Sea basin (after Heybroek and others, 1967)

Fig. 2. Bottom sediment distribution in the North Sea

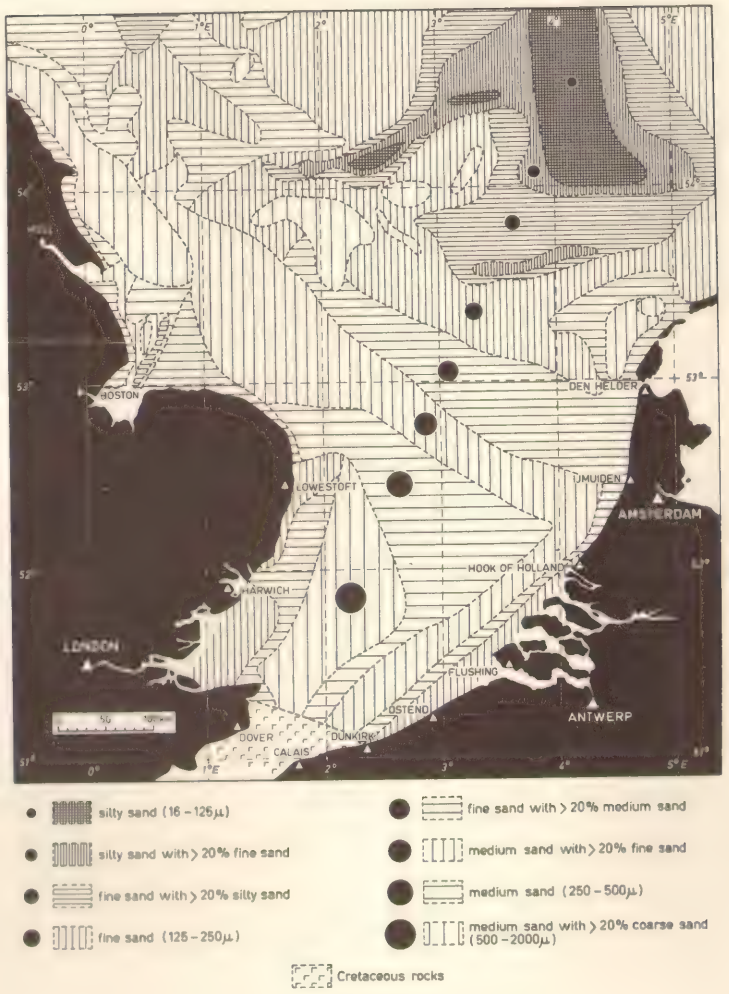


Fig. 3. Grain size of sediments in the southern North Sea (after Jarke, 1956).

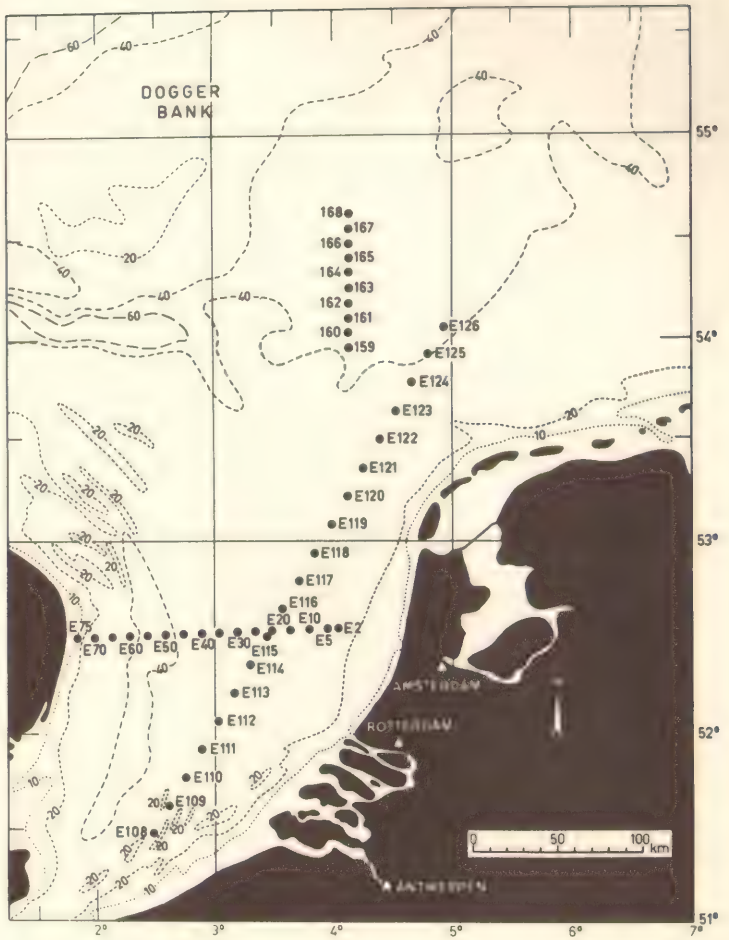
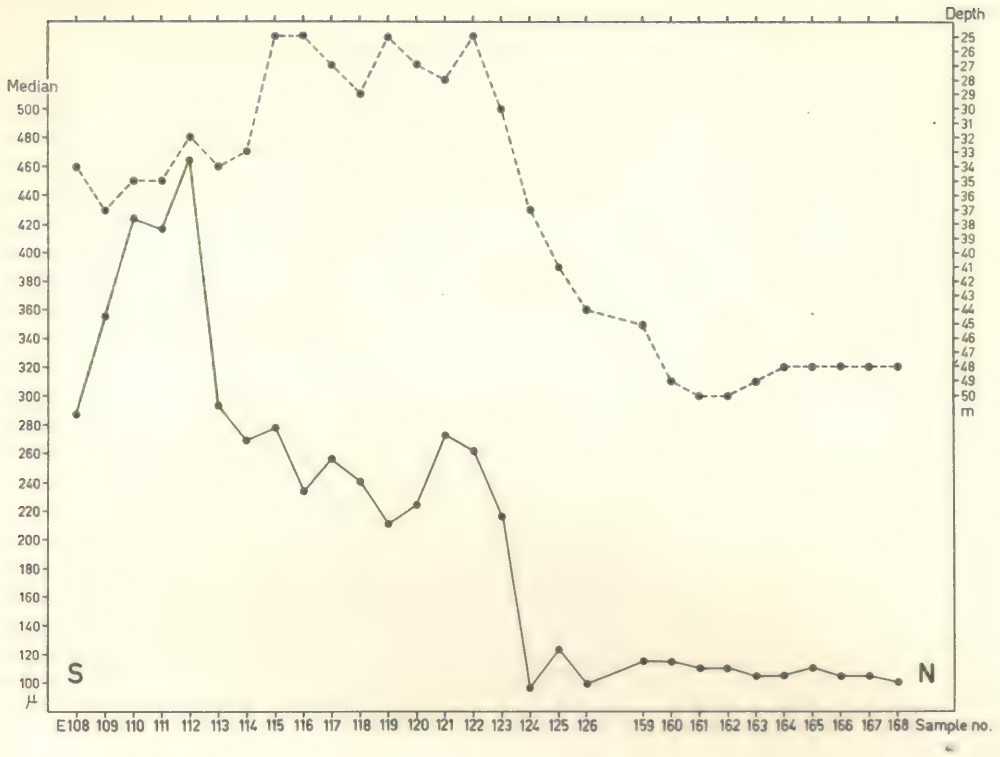


Fig. 4. Location of samples in the Southern Bight, North Sea. The E samples were taken in collaboration with Eisma.



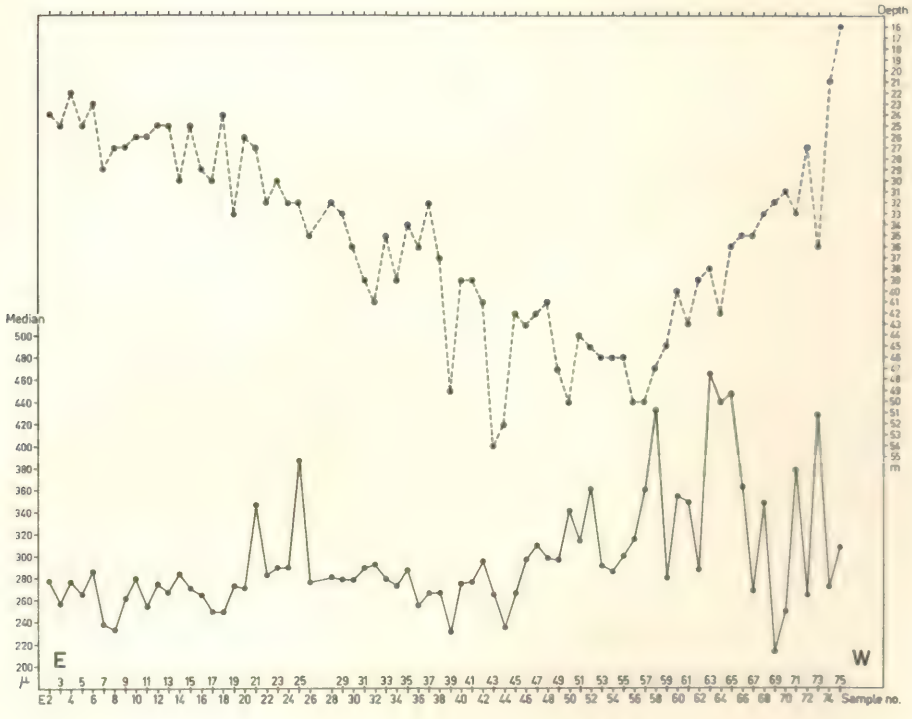


Fig. 5. Median plotted against depth of the samples indicated on Fig. 4, showing increasing fineness towards the north and towards the east.

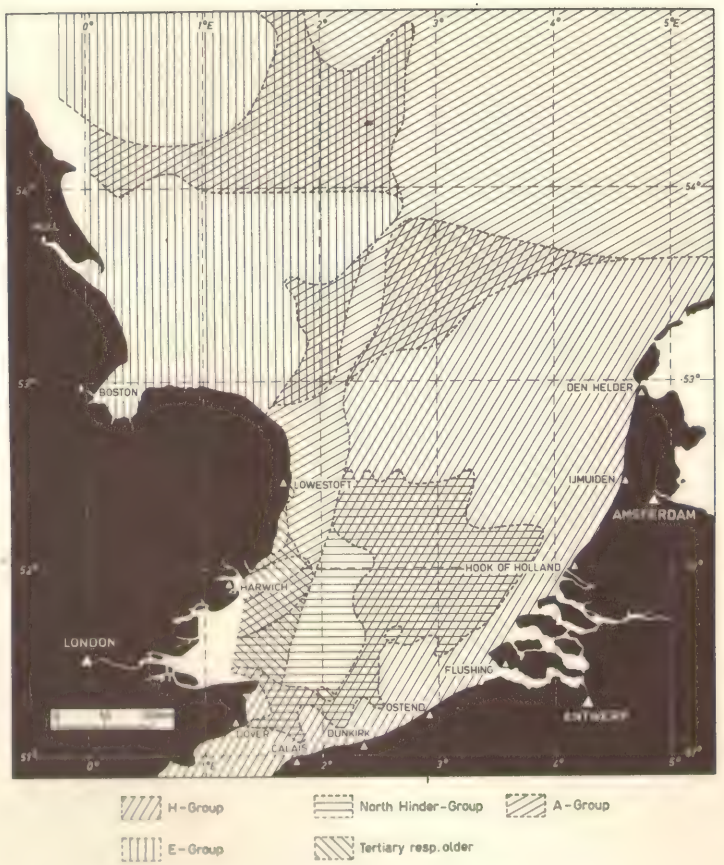


Fig. 6. Heavy mineral assemblages according to Baak (1936). A group: garnet-epidote-hornblende; origin, Scandinavia. H group: garnet-epidote-saussurite-hornblende; origin, Rhine and Scandinavia. E group: garnet-augite-hornblende-epidote; origin, England. North Hinder group: augite-hornblende-saussurite; origin, Rhine. Tertiary group: zircon-rutile-garnet-epidote-hornblende; origin, Tertiary sediments.

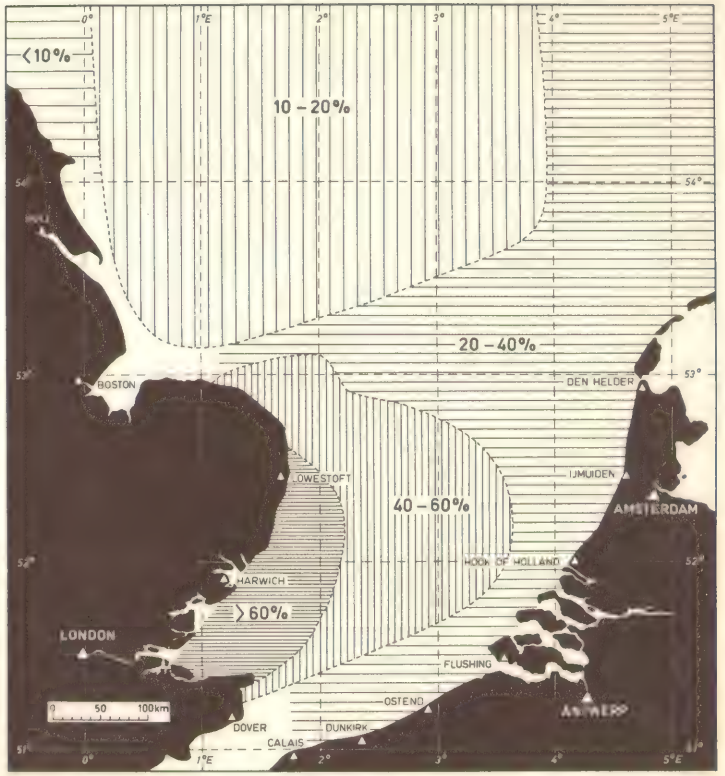


Fig. 7. Percentage of polished quartz grains in North Sea sediments (after Cailleux, 1942)

clays of Eemian and Holsteinian age occur.

The foregoing discussion shows that at present rivers form the chief source of mud, but it is probable that at the beginning of the Holocene eroded boulder clays also supplied considerable quantities of mud.

SAND

The major part of the North Sea floor consists of sand. It appears that the grain size decreases from south to north (Jarke, 1956; Stride, 1963) which might indicate that currents transported the fines towards the north (Fig. 3).

A median of 250-500 μ is found in the Hinder Bank region, a median of 200-300 μ is found off the Dutch west coast and a median of 100-125 μ is found in the Oyster Ground (Figs. 4 and 5).

The origin of the sand in the southern part of the North Sea has been attributed to longshore drift from the Atlantic coasts of France and Britain, to transportation by Pleistocene rivers such as Rhine, Meuse, Thames and Scheldt, and to transportation by ice sheets from Scandinavia. Results obtained by heavy mineral analysis (Fig. 6, Baak, 1936; Kruit, 1963) and roundness studies of quartz grains (Fig. 7, Cailleux, 1942)

Fig. 8. Gravel types of the southern North Sea. The sand wave areas are after Dingle (1965) and Houbolt (1968)

show that the present North Sea floor consists of sand originating chiefly from moraines, fluvio-glacial deposits and cover sands, with the exception of the coastal areas where a mixing with fluvial sands occurred, and the extreme southern part of the North Sea, where Rhine sands were deposited in the deeper areas. With the rise of sea level the gradient of the Rhine decreased and sand transport came to a standstill.

It is probable that since the sea water circulation around the British Isles was restored by the rising sea level in the beginning of the Holocene, fine sand has been transported to the north by the currents. Use of radioactive tracer has confirmed sediment transport parallel to the west coast of Holland (Svasek and Engel, 1961).

An interesting phenomenon is the occurrence of sand waves (megaripples) in the southern North Sea at a water depth of more than 20 m (Fig. 8). They occur in areas with strong tidal currents and were described among others by Van Veen (1936), Cloet (1954), Stride (1963), Dingle (1965), Langeraar (1966) and Houbolt (1968). A height of 10 m and a wave-length of 300 m is quite normal. Even detailed hydrographic surveying during a number of successive years could not prove migration of the sand waves (Langeraar, 1966). In this case the pattern correction of the decca chain used was less than 60 m, which means that a possible displacement of that order was not registered.

Along the Dutch coast the majority of the sand waves strikes north-west to south-east and has asymmetrical cross-sections with steeper northern slopes.

Observations at Smith's Knoll light vessel (Draper, 1966) showed that in water of 100 ft (33 m) depth in 10 per cent of the time particle speeds of 15 cm/sec are reached, which means that sand grains of 250 μ can be taken up in that case. As the median of sand samples taken between Holland and England is always coarser than 200 μ, this means that during storms considerable quantities of sand are transported that may influence the shape of the sand waves. Probably the sand transport occurs in the form of smaller ripples moving over the slopes of the megaripples.

Sections across sand waves show that in some cases coarser sand forms the crest (Houbolt, 1968). Analogous results were obtained by Eisma (1968), who found a difference between the average of medians of sand in the crests (300 μ) and those in the troughs (270 μ) of 30 μ.

Shape studies by Winkelmolen (1969) of samples taken in the sand wave area off the Dutch coast revealed that the shape distribution of the coarse sand from water deeper than 15 m cannot be attributed completely to the present day hydro-

graphical situation. Therefore the present shape and size distribution of these samples points to higher energy conditions, presumably during the beginning of the post-glacial transgression. Hence, the sand waves may have been shaped initially by more powerful tidal currents than occur at present, that is, in the Lower Holocene when sea level stood 15 m lower.

The same applies to the formation of elongated sand ridges with a length of 20 km or more in the southern North Sea. According to Van Veen (1936) they were formed by strong ebb and flood currents that piled up sand when sea level stood 15 m lower. It appears that, for example, the Hinder Banks off the Belgian coast do not indicate a noticeable shift on hydrographic charts of the last 40 years (Veenstra, 1964). According to Houbolt (1968) the ridges in the Well Bank area off the British coast move slowly northeastwards. He could also show that in the southern North Sea a distinction has to be made between erosional remnants, such as the Brown Ridge, that have a core of Pleistocene sediments, and accumulation forms to which the Flemish Bank, Hinder Banks and Well Banks, belong. The latter type should consist partly of Rhine sands deposited before the postglacial transgression reworked these sediments.

GRAVEL

The gravel of the North Sea does not form a single unit in distribution, composition and origin. The gravel content of the samples varies between 0.5 and 60 per cent, the rest being sand. There are at least seven possible origins of the gravel.

1. The gravel is transported at the present time by wave erosion from the English coast or from submarine outcrops. This hypothesis is rejected, because the present currents cannot move the pebbles. The maximal currents at the surface do not exceed a velocity of 1.5 m/sec, and in most places do not even reach 0.5 m/sec. The bottom currents will be weaker.

2. The gravel was supplied by Tertiary and Pleistocene rivers such as Rhine, Meuse, Scheldt and Thames. Although rivers brought large quantities of sand and gravel into the North Sea during the Pleistocene, it is improbable that gravels of Rhine and Meuse outcrop on the sea floor. In a 50 km wide zone parallel to the Dutch coast the sea is shallower than 30 m. At the Dutch coast fluvial gravel is situated at 25 m below sea level (Pons, 1954) and the depth increases in off shore direction. Moreover, sparker sections made by Houbolt (1968) show that the Holocene sands in the southern part of the North Sea vary in thickness from 5 to 10 m. Therefore most North Sea gravels are not of fluvial origin. Exceptions are the gravels found near the south-eastern coast of Britain in 20 m of water and off the Belgian coast in water 35 m deep. The first gravel type might belong to ancient Thames deposits, whereas the second type could have been supplied by the river Scheldt.

3. The gravel could represent a basal conglomerate formed during post-glacial rise of sea level. As the depth of the gravel varies between 20 and 50 m, this is improbable.

4. The gravel forms a part of lag deposits from Pleistocene moraines. This origin must be discarded for the Southern Bight area, as this part of the North Sea has never been glaciated.

5. The gravel has been deposited by meltwater or ice-floes of the Pleistocene ice sheets. As only very few igneous rocks are found in the gravels of the Southern Bight, this origin must be rejected for this area.

6. The gravel belongs to a weathering residue that once covered the Ardennes and the North Downs, formed during the Tertiary. It is possible that part of the gravel in the Southern Bight has this origin.

7. The gravel forms a continuation of alluvial terraces or submerged pebble beaches. It is probable that in and near the entrance of the Thames Estuary gravels of this group are found.

From the foregoing discussion it appears that there can be no common origin for all the gravel. Hence, several types must be distinguished (Veenstra, 1969). Based on composition the following four types (Fig. 8) have been recognised.

Flint-limestone-quartzite

This type is found in the area between Belgium and England. In the neighbourhood of the Hinder Banks about 65 per cent of the gravel is coarser than 8 mm, the depth of the water is about 35 m. In the fraction 8 - 16 mm about 10 per cent of the pebbles are well-rounded and 25 per cent angular (Table 1). Igneous rocks (porphyries, granite) form less than 5 per cent of the pebbles. Other pebbles consist of chalk, oolite (Jurassic), brown flint (Cenomanian), schists (Devonian) and mosaic quartz originating from a metamorphic region. Presumably the gravel was transported by Tertiary rivers that took up the resistant materials from a weathering residue on the Ardennes.

The gravels found near the entrance of the Thames Estuary at a depth of about 20 m contain much flint. About 11 per cent is well-rounded and pebbles with percussion marks occur. The average percentage of angular flint fragments is 18 per cent. The gravel, 90 per cent of which comprises material coarser than 8 mm, presumably represents submerged beach material.

TABLE 1

area	sample no.	size fraction 8-16 mm R	SA	A	weight percentage · 8 mm
Hinder Banks	79	3	71	26	80.2
	82	15	63	22	50.1
	92	12	60	28	66.5
	110	9	66	25	76.6
Thames Estuary	HA	9	70	21	89.9
	TH	13	74	13	91.4
	SH	11	70	19	87.4
Yorkshire Coast	G2	19	78	3	25.2
	G3	9	81	10	31.7
	F2	6	79	15	61.7
Inner Well Bank Rough	137	16	68	16	77.1
	145	4	74	22	27.1
	247	6	78	16	32.1
	A2	3	83	14	34.6
	B2	5	83	12	24.1
Dogger Bank	228	8	79	13	68.0
	755	10	81	9	53.4
	P5	5	65	30	11.5
Cleaver Bank	123	3	79	18	19.7
	220	8	71	21	70.7
Texel Rough	TX	4	74	22	66.0
	VL	3	73	24	86.5

Roundness distribution of size fraction 8 - 16 mm and weight percentage of fractions coarser than 8 mm.

R, well-rounded pebbles.

SA, subrounded pebbles and angular pebbles with rounded edges or corners.

A, angular pebbles.

Sandstone-limestone-porphyry

This gravel type occurs between Yorkshire and the Dogger Bank at depths of some 30 m. The amount of well-rounded pebbles in the fraction 8 - 16 mm varies between 6 and 19 per cent and the amount of angular pebbles is between 3 and 15 per cent off the Yorkshire coast. The average content of material coarser than 8 mm is 39 per cent, which is almost constant over the whole area between Yorkshire and the Dogger Bank.

Further to the east on the Well Bank Rough the amount of well-rounded pebbles varies between 3 and 16 per cent and that of angular pebbles between 12 and 22 per cent. The average pebble content coarser than 8 mm is 39 per cent. The water depth is about 34 m.

On the Dogger Bank well-rounded pebbles vary between 5 and 10 per cent and angular pebbles between 9 and 30 per cent. The water depth of the samples is about 34 m and 44 per cent of the gravel is coarser than 8 mm. The conclusion is that the gravel near the British coast is better rounded. Encrusting organisms such as bryozoans and serpulids on the pebbles show that the gravel is stationary at present.

The gravels of the Well Bank Rough and the Dogger Bank are more angular, because they contain more flint which is angular in most cases, whereas the other material is of the same roundness in the whole area.

The composition of the gravels shows Jurassic oolites, Cretaceous flint, Carboniferous rocks such as sandstone with coal fragments, crinoidal limestone, and purple, brown and black porphyries with rectangular phenocrysts, probably derived from porphyritic rocks in the Cheviots (Eastwood, 1953).

The occurrence of these rocks together points to a large source area. As Weichselian boulder clay in Durham possesses the same petrographic composition, the gravels in the adjacent part of the North Sea are attributed to lag deposits of moraines built up by a Scottish glacier.

South of the Dogger Bank boulder clay that shows a profusion of Carboniferous and Jurassic spores and therefore differs in composition from the Dutch moraines of Saale age (Zagwijn and Veenstra, 1966) is found on the sea floor.

Porphyry-granite-flint

This gravel type is found on the Cleaver Bank, south-east of the Dogger Bank. The average water depth of the samples is 36 m. This gravel differs from the previous type in having a higher flint content, but lower limestone and sandstone contents. From the composition it seems to be a Scandinavian gravel with an admixture of British rock fragments. Owing to the higher flint content this gravel is more angular than the previous type, and the percentage of well-rounded pebbles varies between 3 and 8 and the percentage of angular pebbles between 18 and 21. The average gravel content coarser than 8 mm is 45 per cent.

Quartzite-flint-granite

This gravel type occurs at the Texel Rough near the Dutch coast at a water depth of 25 m. Here pebbles of vein quartz and sandstone have also been included under the heading 'quartzite'. Gravel of similar composition in the Vlie tidal inlet is derived from boulder clay layers that were eroded by strong tidal currents. The boulder clay is overlain by Eem clay and belongs to the Penultimate or Saale Glaciation.

The gravel contains pebbles of black flint and red feldspar, which points to a Scandinavian origin. On the whole, it is poorly rounded. Only 4 per cent is well-rounded and 23 per cent is angular (Table 1). About 75 per cent of the gravel is coarser than 8 mm.

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Late Quaternary geology of the North Sea south-east of the Dogger Bank

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SUMMARY

The Netherlands Geological Survey is investigating the geology of the Dutch part of the North Sea. The sampling equipment consists of a small vibracorer and a big one; the latter can be used to make deeper straight-flush borings in unconsolidated material. Additional information is gathered from samples collected by oil companies. A survey with a sonic profiler has been carried out.

The geology as published earlier is reviewed. New data are given, of which the most important deal with the presence of the interglacial Eemian deposits and the build-up of the Dogger Bank. As to the first, Eemian sediments appeared to be present in a broad valley, reaching to a depth of over 80 m below mean sea level. Concerning the Dogger Bank it is stated that the author's view held earlier must be discarded. Glaciogenic sediments do not occur at shallower depths here, but are found at the same depth as on the flanks. It is assumed that on top of these sediments Weichselian meltwater sands will be present, while later finer fluvioglacial deposits and cover sands have been deposited.

SOMMAIRE

Le Service Géologique des Pays-Bas étudie actuellement la géologie de la partie hollandaise de la Mer du Nord. Les échantillons sont pris per deux 'vibracorers', l'un petit et l'autre grand. Celui-ci peut se servir pour faire des sondages 'straight-flush' plus profonds dans le matérial non consolidé. Des informations supplémentaires sont dérivées des échantillons obtenus par l'industrie pétrolière. Un levé avec un sondeur sonore a été effectué.

La géologie déjà publiée est passée en revue. Entre les données nouvelles, qui sont fournies, les plus importantes traitent de la présence des dépôts interglaciaires Eemiens et de l'accroissement du Dogger-Bank. Quant au premier, des sédiments Eemiens semblaient être présents dans une vallée large, atteignant une profondeur de plus de 80 m au dessous du niveau moyen de la mer. En ce qui concerne le Dogger-Bank, on déclare que l'avis antérieur de l'auteur doit être abandonné. Des sédiments glaciogéniques ne se présentent pas ici aux niveaux moins profonds, mais se trouve au même niveau que sur les flancs. On suppose que, par dessus de ces sédiments, se trouveront des sables d'eau de forte Weichselienne, tandis que des dépôts fluvioglaciaires plus fins, et un manteau de sable ont été deposés plus tard.

ZUSAMMENFASSUNG

Das niederländische geologische Amt untersucht die Geologie des holländischen Teils der Nordsee. Die Mustersammlungsgerätschaft besteht aus einem kleinen und einem grossen 'Vibracorer'; man kann diesen benutzen, um tiefere 'straight-flush' Bohrungen in unkonsolidiertem Stoff zu machen. Man sammelt weitere Information von Mustern, die von Ölgesellschaften gesammelt sind. Eine genaue Untersuchung mit einem Schallmessgerät wird durchgeführt.

Man wiederholt die Geologie, die früher veröffentlicht war. Man gibt Tatsachen, aus welchen die wichtigsten des Vorkommen von Interglazial-Eemianablagerungen und den Aufbau der Doggerbank behandeln. In Betreff der ersten, scheinen Eemiansedimente in einem breiten Teil anwesend zu sein. Sie reichen bis einer Tiefe von mehr als 80 m unter mittlerem Meeresniveau. In Bezug auf der Dogger Bank, meldet man, dass die Meinung des Verfassers, die er früher behauptet, abgelegt sein muss. Sedimente von einem Gletscherursprung befinden sich nicht in seichteren Tiefen hier, sondern befinden sich in der gleichen Tiefe als zu den Seiten. Man nimmt es an, dass oben auf diesen Sedimenten befinden sich Weichsel-Schmelzwasser, während später Fluviatilgletscherablagerungen und Decksände abgelagert worden sind.

Late Quaternary geology of the North Sea south-east of the Dogger Bank

E. OELE

INTRODUCTION

After a few limited investigations at sea the Netherlands Geological Survey some years ago started the mapping of the Dutch part of the North Sea. Co-operation with the North Sea Department of the Department of Harbour Entrances of Rijkswaterstaat (Ministry of Public Works) provided the Survey with the collection of samples. It has already resulted in the publication of a geological map of the northern half of the Dutch North Sea.

Recently gathered new data have yielded information that make a preliminary publication worth while.

COLLECTION OF DATA

Various sources of information on the subbottom sediments are available. Partly they consist of the samples collected by the Geological Survey with its own equipment, partly of data put at the disposal of the Geological Survey by the Ministry of Public Works and by oil companies.

Conditions in the North Sea are such that special equipment is required. Various kinds of equipment, o. a. gravity corers and penetrating echo sounders yielding good results in other areas, did not give any results in the North Sea.

Sampling Equipment of the Geological Survey

Two types of vibracorers are used by the Geological Survey. A small corer, developed by the B. I. P. M., makes it possible to collect undisturbed samples with a length of 4 m. A special vibracorer, the Geodoff, designed by the Geological Survey and Conrad Stork N. V., has a sampling length of 6 m (undisturbed samples). Apart from its greater sampling length, the advantage of the Geodoff is that the corer itself lifts the sampling tube out of the seabottom. A drawback of both corers is their restricted penetration in tough clays.

The use of the technical qualities of the Geodoff enabled us to make straight-flush borings, and attain deeper penetration (over 10 m) independent of the occurrence of clay beds.

Data Gathered by Oil Companies

A great number of oil companies holding exploration licences for the Dutch part of the North Sea are investigating the subbottom conditions of the future rig-sites. Most of the companies had borings carried out to depths of 20 to 30 m below sea bottom. The results of the investigations and the samples are placed at the disposal of the Geological Survey. Various companies gave the Geological Survey permission to publish the data and their geological interpretation.

Continuous Subbottom Profiler

On behalf of the above mentioned North Sea Department of Waterworks a survey has been carried out by C. E. S. C. O. (The Hague) with a continuous sonic profiler (Sonia). The frequency of the pulse transmitted is 3 kc/s. In the North Sea the penetration appeared to be 30 m on an average. In some instances a penetration of 50 m was recorded. To a certain extent gravel, sand and clay can be discerned. The records (or sonograms) of this survey, which covered with its 1200 miles the whole Dutch part of the North Sea, have been geologically interpreted by the Geological Survey.

LATE QUATERNARY GEOLOGY SOUTH-EAST OF THE DOGGER BANK AS PRESENTED BEFORE

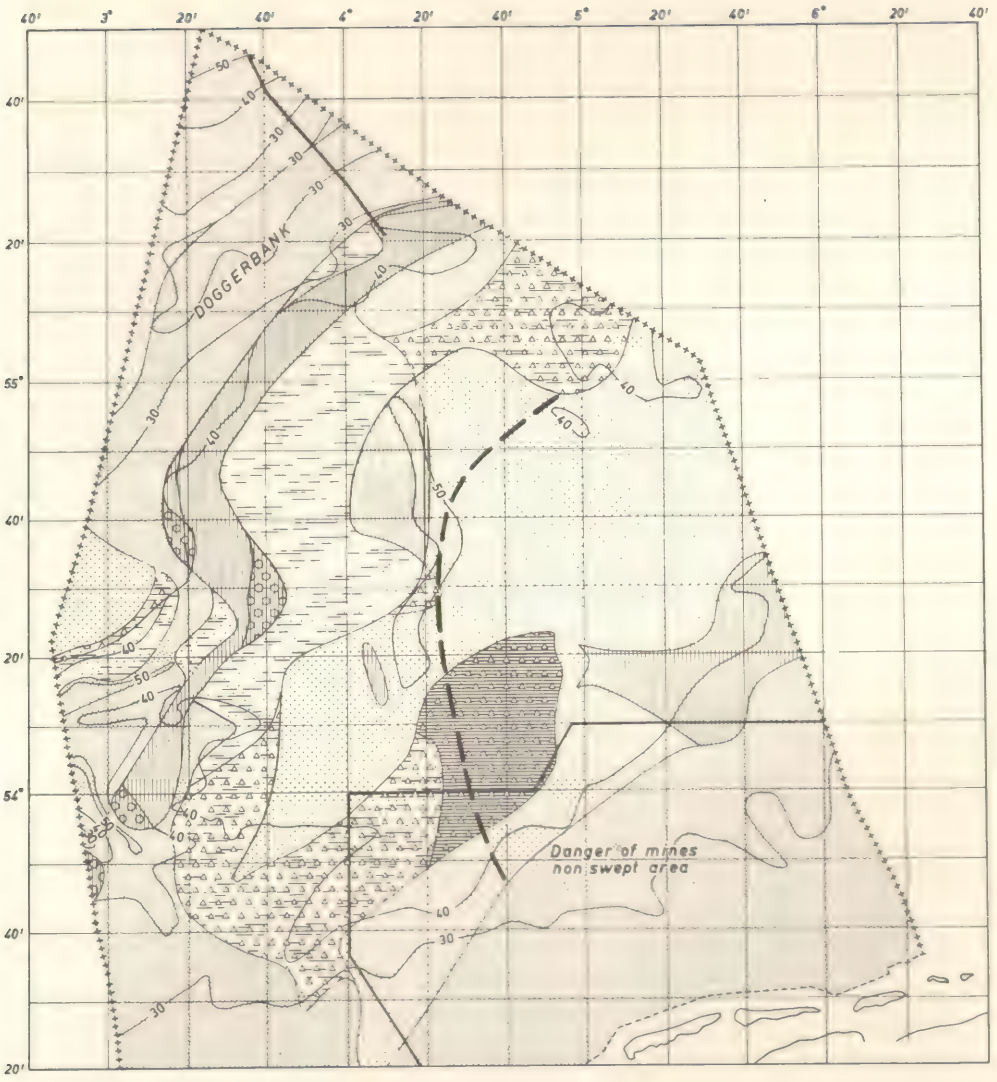
Based only on samples collected with the small vibracorer, a map was drawn, showing the Quaternary geology of the northern half of the Dutch part of the North Sea (Oele, 1969). The following series of events was presented (Fig. 1).

Elsterian Glacial

The oldest deposits encountered consist of glacial meltwater deposits, mostly clay, occasionally silt. The clay is assumed to underlie the younger sediments throughout the whole area. Its fluvioglacial origin was deduced from its pollen content. In a few instances varve-like structures were observed.

Saalian Glacial

After the Holsteinian Interglacial, of which period no deposits were found, the Saalian glaciers moulded the surface of the above mentioned fluvioglacial clays. A broad valley, in which the clay crops out, was created by glacial scour. This was deduced from the clay surface which slopes from the Dutch coastal area into the depression, south-east of the Dogger Bank, and rises again at the flank of the Dogger Bank. Extrapolation suggested that the Dogger Bank might be considered as an erosion remainder of the



LEGEND

HOLOCENE

Dunkirk deposits
Young seasand more than 2m in thickness on fluvioglacial deposits; intercalations possible

Elbow deposits
Early holocene brackisch fine sands and clays

Idem, on Twente Formation

Lower peat
Lower peat deposits on Twente Formation

PLEISTOCENE - Twente Formation
Coversand

Drente Formation
Boulder clay

Peelo Formation
Fluvioglacial deposits, mainly clay

Note: All profile-types mentioned are supposed to overlie fluvioglacial deposits and to underlie young seasand less than 2m in thickness

W Boundery of Eemian deposits in glacial valley

Fig. 1. Late Quaternary geology of the northern half of the Dutch part of the North Sea (modified after Oele, 1969).

fluvioglacial clay.

The Saalian glaciers deposited boulder clay on the south-east flank of the Dogger Bank.

Eemian Interglacial

No Eemian sediments were found. This can be explained by a rise of sea level so rapid that sedimentation could not keep up with it.

Weichselian Glacial

During the last glaciation the area was not covered by glaciers. Sediments were deposited under periglacial conditions. They consist mainly of cover sands with an intercalation of a humic layer. The sediments are mainly found in the eastern part of the investigated area, where they crop out or are covered by only a thin layer of Holocene deposits. In the early Weichselian as well as during the late Weichselian fluvial clays were deposited.

The Holocene Sequence

As found in the area of investigation it shows a remarkable resemblance to that of western Holland. In the south-eastern part a basal peat bed was deposited on top of the Weichselian sediments. After the peat-growth at first fluvial sedimentation took place again, but gradually it became brackish and marine, while the sediments assumed a more sandy character. The sequence was closed by a layer of Recent to subRecent sea sand. Since the older Holocene deposits named Elbow deposits range in age from pre-Boreal to late Boreal, a stratigraphic hiatus is supposed to exist.

SUPPLEMENTARY DATA

After the geological map just reviewed was published, most valuable supplementary data became available. Although the map was based on a relatively small number of borings, its basic outline appeared to be in accordance with reality. This holds good especially for the boundaries of the various Holocene and Weichselian deposits.

The 'basal' fluvioglacial clay has a thickness of over 30 m on the flank of the Dogger Bank, as was established by borings made on behalf of B.P., the base of the clay occurring at a depth of approximately 70 m below mean sea level. Fine sands underlie the clay. Lack of fossils makes a dating of this sand impossible at the moment. The grain size as well as the intercalation of clay laminae might indicate that the sands were deposited under conditions similar to those during the formation of the Weichselian cover sands. It is noted that the plane dividing the fluvioglacial clay and the sands coincides with a horizon which can be traced throughout the area of investigation with only minor fluctuations in its depth.

The Saalian boulder clay was found to crop out as far south as Botney Cut (approximately 54°N 3°E), thus stretching further southward than shown on the map. The Botney Cut depression with its present depth of over 70 m represents a valley cut into the glacial deposits and later partly filled up with younger sediments, but its origin is not yet clear. A fluviatile erosion during the Weichselian seems doubtful. During the major part of this glacial a periglacial climate governed the area, while the capacity of the streams during the rest of the Weichselian was probably too low to account for the formation of such a broad and deep valley in the tough glacial clays.

The Holocene basal peat was sampled here as well as on the south-eastern flank of the Dogger Bank. Borings made on behalf of B.P. also found Holocene basal peat.

A radiocarbon dating of two peat samples earlier collected and considered to represent the basal peat in blocks F14 and F15 yielded ages of 9730 and 9880 years B.P. respectively. The figure is fully in accordance with the curve of the post-glacial sea-level rise drawn by Jelgersma (1961), when the curve is extrapolated to the pre-Boreal, because the peat occurs at 47 and 46 m respectively below mean sea level.

The most remarkable discovery, however, is the presence of Eemian Interglacial sediments. These sediments constitute the infilling material of a broad valley cut in the glacial deposits. Samples provided by Placid International demonstrate that at a depth of more than 80 m below mean sea level, that is, well below the base of the fluvioglacial clay, the valley base was not yet reached.

The lower part of the Eemian deposits consists of fine sands, which probably settled in a littoral environment, considering the scarcity of fossils. In this series deeply incised gullies (Fig. 2), were filled up with sands, which represent brackish environment, as was concluded by Van Voorthuysen from their foraminiferal content. An interrupted clay bed marks the end of this sedimentation. During the Late Eemian or perhaps even the early Weischselian a series of fluviatile sands was deposited on top of the older Eemian sediments. The course of the boundaries of the valley is only roughly known. The similarity of infilling material suggests that the valley is related to the glacial meltwater valley of the Hunze in the Northern Netherlands as described by Ter Wee (1963).

New data became available on the Dogger Bank too. Fig. 3 shows a cross-section over

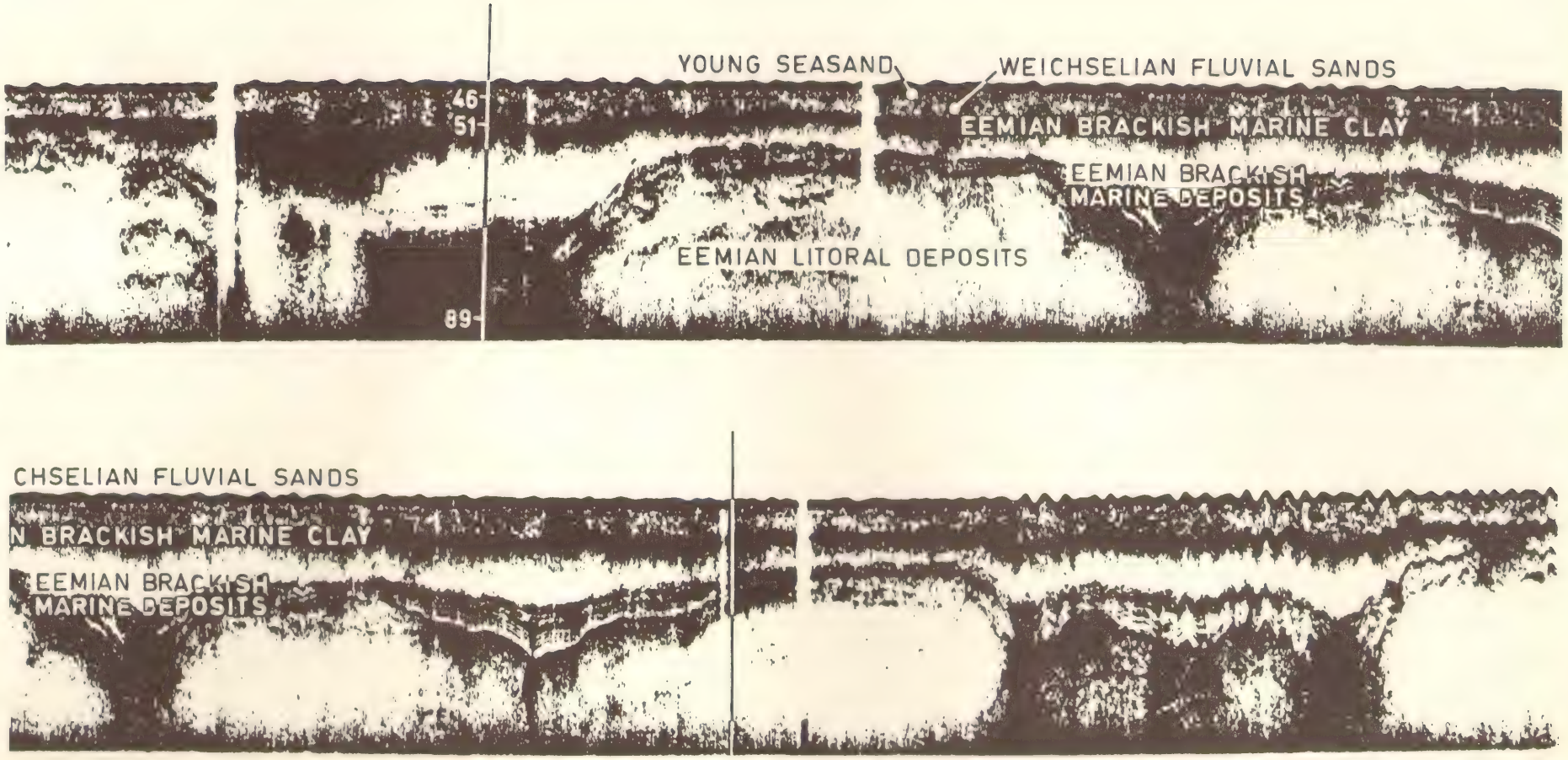


Fig. 2. 'Sonia' record of Interglacial Eemian infilling of glacial valley. Distance between plottings approximately 1900 m. Right hand end of upper section overlaps left hand end of lower section.

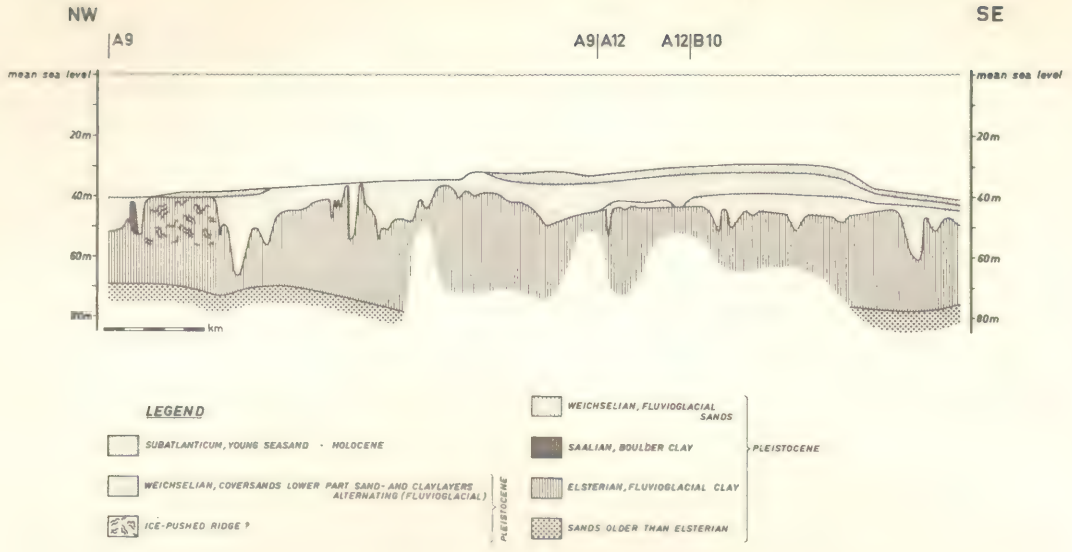


Fig. 3. Interpreted 'Sonia' section over the Dogger Bank. For location see Fig. 1.

the Dogger Bank drawn after the records of Cesco's 'Sonia'. The records demonstrate that the surface of the fluvioglacial clay rises from the depression south-east of the Dogger Bank in the direction of this bank. This had already been concluded from the borings made before, which then led to the suggestion that a continued rise might explain the existence of the Dogger Bank. As is visualised on Fig. 3 the surface of the clay, however, remains more or less horizontal under the Dogger Bank – in other places it even proved to form a shallow depression – to rise again on the north-west flank of the Dogger Bank. It is possible that the upper part of the clay actually consists of boulder clay, which can hardly be distinguished from the fluvioglacial clay by means of 'Sonia' equipment.

On top of the glaciogenic deposits Weichselian sands are thought to be present, from the fact that the young sea sand consists of reworked cover sands and from the presence of mammalian bones (compare Stride, 1959). As can be seen in Fig. 3 north of the Dogger Bank an ice-pushed ridge seems to be present. This would be in accordance with the idea of Valentin (1955), who holds this to be the southerly limit of the Weichsel glaciers. The data presented here causes my earlier views on the Dogger Bank to be discarded. The elevation is not the result of a glacial high. I would suggest the following events. In front of the Weichselian glaciers proglacial deposits settled, on top of which a series of cover sands with clayey layers was deposited after the ice had started to retreat. A layer of young sea sand closed the sequence. Deep valleys were cut in the glacial deposits and filled up with younger material.

CONCLUDING REMARKS

I still think that during the Elsterian glaciation only the fluvioglacial meltwater clay was deposited. Its surface was moulded during the penultimate glaciation, the Saalian, with subsequent deposition of boulder clay. During the last glaciation, the Weichselian, the area was not covered by glaciers, but belonged to the periglacial zone where cover sands could occur.

The above mentioned data show that the Late Quaternary geology of the investigated area was entirely glacially controlled. The climatic changes, causing extreme fluctuations in the sea level governed the sedimentation pattern. It is remarkable that a comparatively thin series of marine sediments was deposited, despite the occurrence of three transgressions since the deposition of the Elsterian fluvioglacial clay.

Marine sediments of the Holsteinian Interglacial have not been found, which means they have a restricted distribution, if they are present at all, while the distribution of the Eemian Interglacial deposits is restricted to a glacial valley. The Holocene transgression indirectly caused the peat growth and more directly the deposition of the early Holocene Elbow deposits, the latter having a very limited thickness.

The Holocene as well as the Eemian transgression are attended with the formation of gullies, later filled with brackish sediments.

The Holocene sedimentation is marked by a hiatus from early Atlantic to sub-Atlantic times.

ACKNOWLEDGEMENTS

The sonic survey, made on behalf of the North Sea Department of Rijkswaterstaat has led to a much better understanding of the geology of the area of investigation. I am grateful to Ir. Van Dixhoorn, Head of the Department of Harbour Entrances and the North Sea Department for his permission to publish some results of the survey.

Most valuable information has become available from the samples, placed at our disposal by British Petroleum Exploratie Maatschappij Nederland N.V. and Placid International Oil Ltd. Therefore, sincere thanks are due to both oil companies.

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Results of a sparker reconnaissance survey of the eastern and central English Channel

J. Robert

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Results of a sparker reconnaissance survey of the eastern and central English Channel

J. ROBERT.

In the spring and summer of 1969 the Centre National pour l'Exploitation des Océans (CNEXO) organised a coordinated geological reconnaissance survey in the Eastern and Central Channel. With the agreement of the British authorities, the survey was extended up to the English coast from Dungeness to Portland. The limits on the French side were the Franco-Belgian frontier and the Cap de la Hague.

The survey consisted of three different operations: firstly, 2500 km of Etinceleur (high definition sparker) lines, mainly following the hyperbolae of Decca chains 1 and 5, were recorded by a service company. About 1500 km of lines had been already recorded since 1966 in restricted areas off Dunkerque, Calais, le Havre and Cherbourg, and test lines had been run on the Varne, the Colbart, the Bassurelle, the Bassure de Baas and the eastern extremity of the Hurd Deep (Robert, 1969). The interpretation of the records based on analysis of their seismic character allowed differentiation of the homogeneous formations from a seismic standpoint which, by correlations with the coastal outcrops, taking into account the data already published (Dangeard, 1928; King, 1954; Arkell, 1953) gave a rough 'seismic stratigraphy'. The present paper presents the interpretation of sparker records based on the available geological data and the lithological determination of a few samples. Secondly, 1030 bottom coring operations were carried out with a Stetson-Hill Corer along the sparker lines by the service company ship and crew under the control of scientists of the Universities of Caen, Rennes, Rouen and Paris. The aim was to tie the seismic sequences precisely to the stratigraphy. Thirdly, sixty large cores were taken with a 10 in diameter vibrocorer developed by the Institut Français du Pétrole (I. F. P.), and operated from the I. F. P. flexo-drilling ship 'Terebel'. The object was to sample the uppermost 25 ft of sediments along the lines which had shown thick enough sections of recent sediments.

The interpretation of the sparker records led to the separation of four tectonic zones: the southern part of the North Sea; the Boulonnais – Weald anticline; the Dieppe basin; the central part of the Channel.

THE SOUTHERN PART OF THE NORTH SEA

From the tectonic standpoint the southern part of the North Sea basin is characterised by a slight northward regional dip. Its southern limit corresponds to the Cretaceous—Jurassic contact along a Folkestone – Cap Blanc-Nez line. Analysis of the records suggests separation of the Albian-Cenomanian formations slightly unconformable on the Wealden formation of the Boulonnais anticline. Northward, there appear, conformable on the Cenomanian, firstly Turonian and then Senonian, which stretch widely between France and England. Toward the northeast, below the Flandres Banks, a Lower Tertiary section exists conformable with the Upper Cretaceous; this zone corresponds to the western part of the Ypres basin, with Landenian and Ypresian series. Locally off Calais, this basin is limited by a northeast-southwest flexure passing into a fault on Sangatte beach.

THE BOULONNAIS-WEALD ANTICLINAL ZONE

The Boulonnais anticlinal zone continues westward in the Channel and the Weald, the anticlinal axis plunging westnorthwest. The northern limit of this zone had been precisely mapped as a result of the tunnel-drilling programmes. It appears on the sparker records as a flexure underlining the limit of the Chalk outcrops along a line from Cape Blanc-Nez to Folkestone. The southern side of the anticlinal zone consists of the Cretaceous sequence dipping evenly towards the southwest, the Chalk cropping out along a line from the Canche estuary to Eastbourne. The observations of the reflection characters allowed us to distinguish two major sequences: the sand-clay Wealden formations, in which a more detailed analysis would probably show three different members, and the limestone Upper Jurassic formations in which thin layers of Portlandian age (Hallez, 1899) contrast with thick beds of Kimmeridgian. A sequence of anticlines trending westnorthwest in which the Jurassic outcrops as far as the centre of the Dover Strait separates synclines in which the Wealden almost reaches the French coast.

THE DIEPPE BASIN

The term 'Dieppe basin' is proposed as a provisional name for a large area located off Dieppe in which the Chalk and Tertiary rocks outcrop in a regular concentric pattern – a disposition similar to that of the Paris Basin. As no sparker lines have been recorded in a large area off Sussex the relations between this structure and the Tertiary basin of the Isle of Wight is not as yet precisely known.

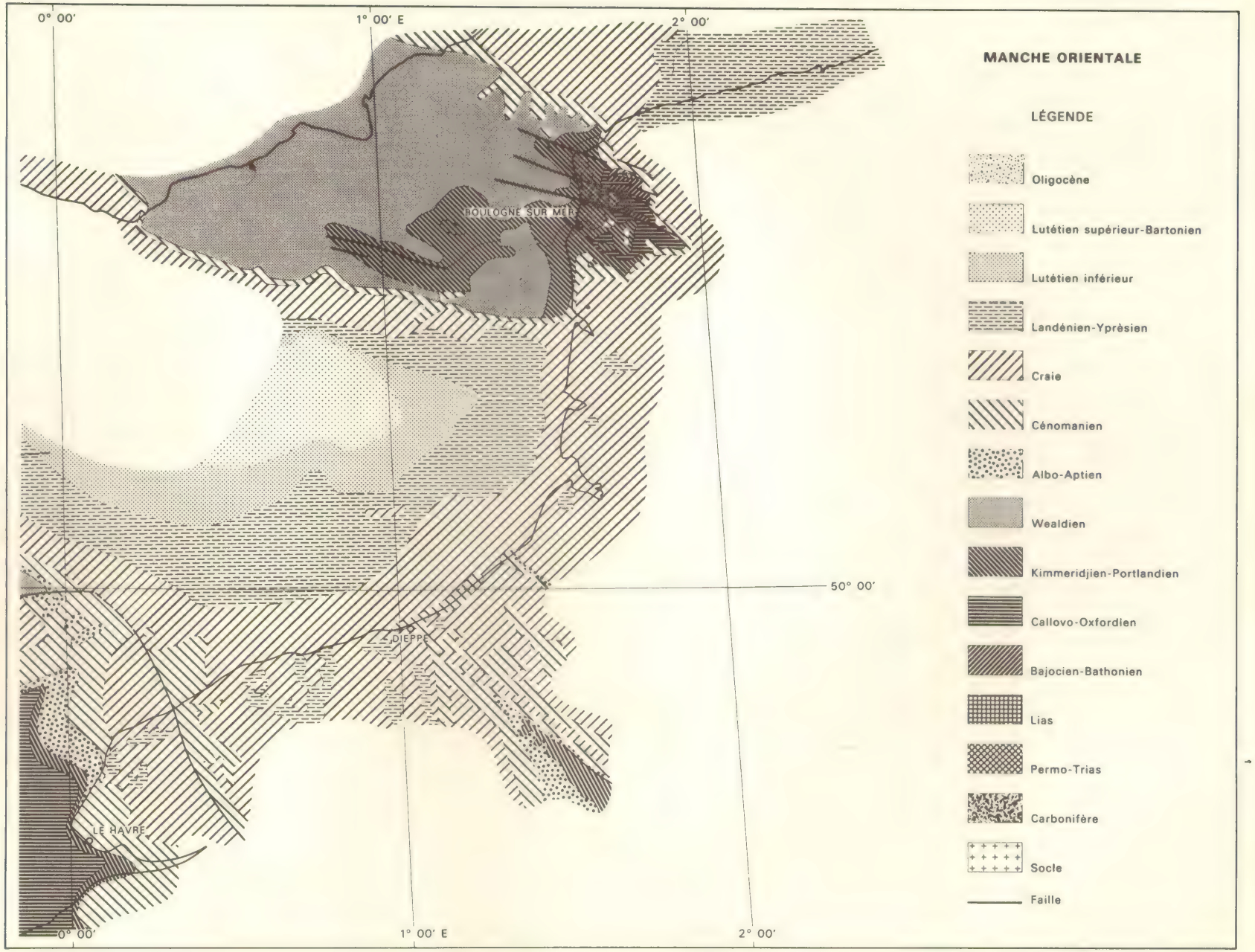

Fig. 1. Geology of the eastern English Channel

Nevertheless, the northern Chalk strip is assumed to be fairly regular and roughly limited by the line joining Eastbourne to the Canche estuary. The Tertiary series are most likely to crop out continuously from off Dieppe to the Isle of Wight in a much larger area than was previously thought (Dangeard, 1928), with restricted zones of Quaternary sedimentation along both coasts.

Considering the stratigraphical sequence along the French coast, the Tertiary is assumed to begin with the Landenian series. Above appears a formation related to the Ypresian and the Lutetian. Though only samples of Lutetian age have been dredged (Dangeard, 1928), the thickness of the Tertiary section (approximately 900 ft) leads us to assume that the sequence may include more recent strata, such as Lower Bartonian sand and limestones, and even Stampian sand. These two formations, which lie flat in the depressed centre of the basin, seen to be covered by recent sediments 6 to 12 ft thick, through which the hardest Lutetian limestone outcrops.

In addition to the noticeable thickening of the Tertiary section with respect to that of the French coastal region, a study of the sparker records shows that it is not possible to trace any seismically homogeneous Tertiary member from one side of the basin to the other, some pinch out westward and others eastward below the basin. This phenomenon may correspond to the imbrication of the Tertiary of English facies already known between the Cotentin and the Isle of Wight (King, 1954; Curry, 1962; Dingwall, 1969), and the facies of the Paris Basin.

The sparker records have not revealed any faults; some slight undulation appears locally but no large feature has been detected. In particular the Pays de Bray anticline, a major structure of the northwestern part of the Paris Basin, does not extend under the Channel.

THE CENTRAL ENGLISH CHANNEL

The Central English Channel may be considered as limited by the line joining the Cap d'Antifer to the Isle of Wight, and the Cap de la Hague to Portland Bill. It is stratigraphically and tectonically uniform.

Around the Barfleur granite bands of Trias and Lias outcrop (Larsonneur, 1969; Larsonneur and Rioult, 1969). Eastward, in the Baie de la Seine, Middle Jurassic (Callovian, Oxfordian and Kimmeridgian) formations are present, generally below Quaternary deposits though they may outcrop locally (Guyader, 1968; Larsonneur, oral communication, 1969). The sparker records confirm these observations and mark precise limits for the different formations. They have also revealed that the Jurassic sequence is much thicker in the Baie de la Seine than in boreholes drilled for oil prospecting in Normandy (Villequier 101 and Yvetot 101); this thickening appears to be particularly important in the Callovian-Oxfordian part of the section.

Towards the centre of the Channel the whole sequence from Lias to Portlandian outcrops along an eastwest trend and dips gently northward. The various Cretaceous transgressions have spread out unconformably over the Jurassic sequences; thus, on the sparker line recorded between Antifer and the Cap de la Heve, three unconformities are visible: Neocomian on Upper Jurassic; Albian-Aptian on Neocomian or Jurassic; and Cenomanian on Albian-Aptian. To the north of the Baie de la Seine, the Cenomanian lies unconformably on the Upper Jurassic and around the Barfleur granite; the Senonian lies unconformably on the Middle Jurassic and oversteps westwards on to the Palaeozoic (Larsonneur and Rioult, 1969).

Lower Tertiary rocks have been deposited unconformably on top of the Senonian Chalk; because of Quaternary erosion they remain only in two restricted areas where a sequence is present at least up to the Lutetian limestone.

From the structural standpoint the Central English Channel may be considered as a large syncline, the axis of which plunges eastwards about 40 miles south of the English coast line (Dingwall, personal communication, 1969). The south flank of this syncline, gently northward dipping, is affected by two major reversed faults trending east-west which affect the whole stratigraphical section up the more recent Tertiary rocks.

These faults have been crossed by sparker lines which show a central crushed zone with diffraction hyperbolae; the high definition of the records shows clearly that on the southern side of this zone the Tertiary strata have been slightly tilted up, and on its northern side, the Jurassic ones have been bent down. This bending along the fault indicates an anticline on the northern side and a syncline on the southern one.

The northern fault lying in mid-channel between Cherbourg and the Isle of Wight appears as the eastern part of a major break trending east-west which affects the Western English Channel from the Lizard thrust eastward (Donovan and Stride, 1961). It brings Kimmeridgian beds on its north side against Lower Tertiary rocks on its south side, both having been discussed by Curry (1962) and by Larsonneur (1969).

The second fault, located roughly along the parallel 50°N, is very similar as it brings Lias, Middle and Upper Jurassic and Lower Cretaceous

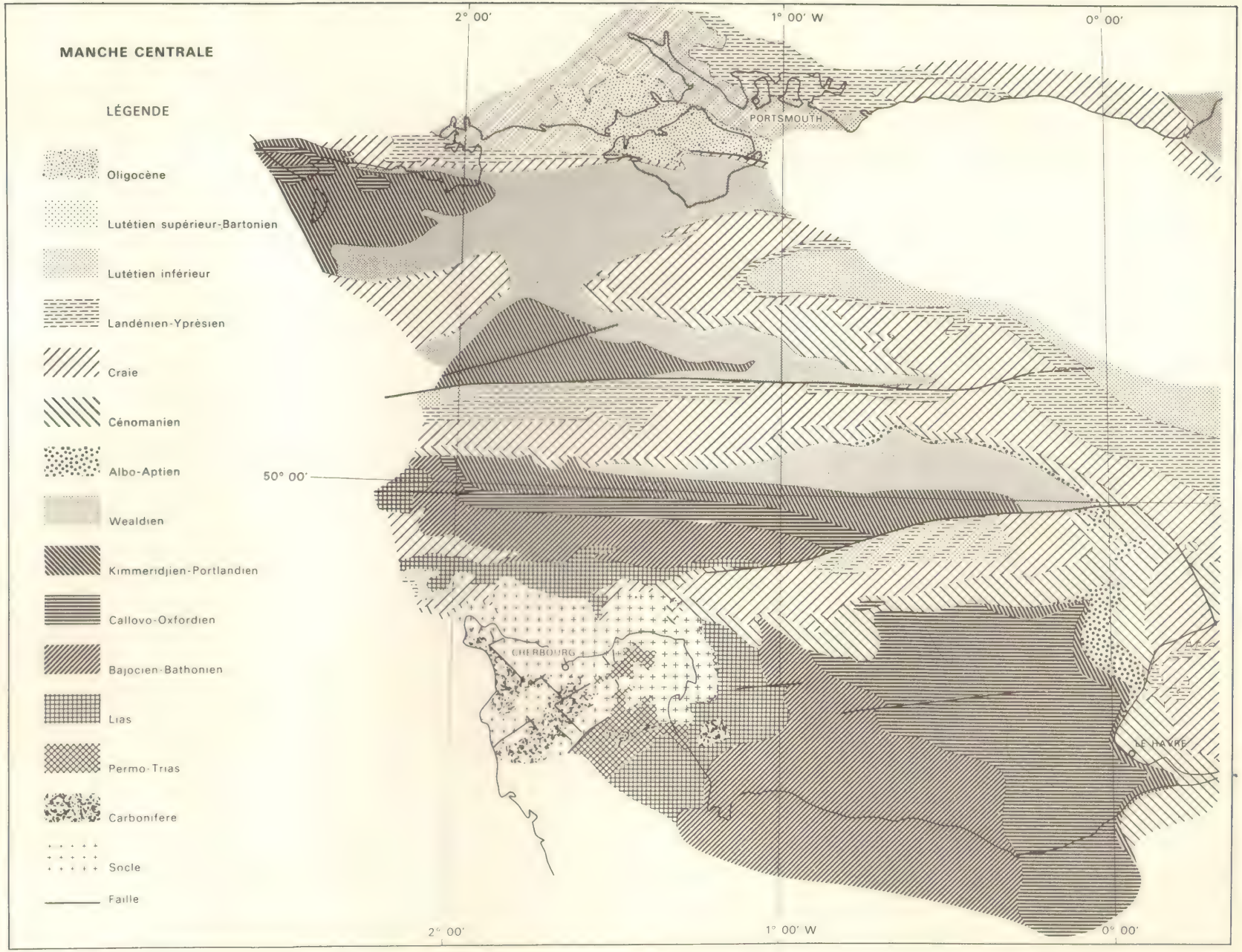

Fig. 2. Geology of the central English Channel

series on the north side against Upper Cretaceous, Lower and Middle Tertiary ones on the south. From the Baie de la Seine to St. Alban's Head, the following outcrop successively: Tertiary in the syncline south of the southern fault; Jurassic in the anticline north of this fault; Cretaceous and Tertiary in the syncline south of the northern fault; Jurassic in the anticline north of this fault; Cretaceous in the main syncline with possible Tertiary relicts; and Jurassic and Cretaceous in the Brixton-Isle of Wight anticline.

CONCLUSIONS

From the interpretation of approximately 4000 km of sparker lines four tectonically and stratigraphically distinct regions have been recognised in the eastern English Channel.

In the southern part of the North Sea there is gently northward dipping Cenomanian and Senonian chalk; the Belgian Lower Tertiary basin does not extend to the west of the western limit of the Flandres Banks.

The northern limit of the Boulonnais-Weald anticline zone is more abrupt than the southern one. The block-faulting tectonics of the Jurassic limestone sequence of the anticline have been attenuated in the plastic sand-clay Wealden formations of the synclines in which they almost reach to the French coast.

In the very flat and tectonically calm Dieppe Basin, fairly thick sequences of Lower and Middle Tertiary, and possibly more recent rock, outcrop widely. Stratigraphical relations between this area and the Tertiary basin of the Isle of Wight are not yet precisely known. The records show lithological members pinching out in opposite directions; this may correspond to rapid facies changes.

In the central English Channel, the progressive extension of the transgressive Cretaceous sequences is clearly visible. This region may be considered as a large east-west syncline, the southern flank of which has been affected by two parallel reverse faults. Each of them has on its south side a syncline in which Tertiary rocks have been preserved and on its north side an anticline which has been eroded down to the Jurassic.

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The continental margin in the Bay of Biscay

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SUMMARY

Geophysical and geological data on this area are reviewed. The main results of the Flexotir seismic reflection surveys made by Comité d'Etudes Marines grouping Institut Français du Pétrole, ELF-ERAP, CFP-TOTAL, SNPA-Aquitaine since 1967 are presented.

Armorican continental margin. The continental shelf appears to be a broad monocline with only Tertiary rocks outcropping, but with older horizons appearing underneath westward. Some horizons can be followed from the edge of the shelf and the slope to the abyssal plain. A subsidence of about 3000 m in the Cretaceous and Tertiary is proved by the present position of a neritic Albian limestone cored on the slope; in addition down faulting of about 1500 m occurred at the end of the Tertiary. These vertical movements can explain the present depth of these layers in the abyssal plain.

Continental margin of Aquitaine. It appears that the Parentis basin is an inland appendage of the Bay of Biscay. From this it can be deduced that the formation of the margin began mainly with vertical movements at the beginning of the Cretaceous continuing up to the end of the Tertiary, accompanied by the sinking of the oceanic area.

The marginal plateau of the Landes is directly linked to the continent. The canyon of Cap Breton is located at the southern edge of a Tertiary subsident block. The Biscay continental slope at the southern side of this area was caused by faults which occurred during the subsidence of this block.

Asturian continental margin. Thick Mesozoic series found on the shelf and on the Asturian marginal plateau show that the Pyrenean area continues on in the northern part of Asturia. The continental slope was formed or was reactivated during the Tertiary.

Abyssal plain. A thick (4000 to 6000 m) linear marginal trough exists from the depression of Cap Ferret to the north-western corner of Spain. It is dissymmetric in section. The greater thickness near the slope is explained by continuous sinking during sedimentation related to faulting of the slope during the Tertiary and perhaps Cretaceous.

The Tertiary and Cretaceous (?) beds of Cantabria and Gascony seamounts pinch out toward the south on top of a rise. These seamounts are tectonic structures caused by Tertiary faults (post-Eocene to Upper Palaeocene).

North of Cantabria there is a large area with less sediments lying on an uneven, broken, surface. In this central area, typical oceanic crust could exist. Along the Armorican slope another linear marginal basin exists with probable piercing structures. These results are compared with magnetic and gravimetric data. It is concluded that Upper Mesozoic must exist in wide areas of the Bay of Biscay and perhaps complete Mesozoic series with possible diapiric structures related to Triassic salt.

The formation of the margin by vertical faulting began between Jurassic and Cretaceous and continued until end of the Tertiary - Pliocene with subsequent sinking of the abyssal plain area and formation of a linear marginal trough at the foot of the continental slope. These results are discussed in regard to the different hypotheses proposed.

SOMMAIRE

Des données géophysiques et géologiques sur cette région sont passées en revue. On présente les résultats les plus importants des études Flexotir par le Moyen de réflexion sismique qu'ont fait depuis 1967 le Comité d'Etudes Marines, un groupement de l'Institut Français du Pétrole, ELF-ERAP, CFP-TOTAL, et SNPA-Aquitaine.

La Marge Continentale Armoricaine. La plate-forme continentale est, semble-t-il, un monoclinal large avec des affleurements seulement de roches Tertiaires. Mais, là dessous, vers l'ouest, des horizons plus anciens se manifestent. On peut suivre certains des horizons à partir du front de la large plate-forme et de la pente, jusqu'à la plaine abyssale. Une subsidence d'environ 3000 m dans le Crétacé et le Tertiaire est prouvée par la position actuelle d'un calcaire néritique albien, dont le noyau se trouve enfoncé sur la pente; en outre une faille d'effondrement d'environ 3000 m s'est produite à la fin du Tertiaire. Ces mouvements verticaux peuvent expliquer la profondeur actuelle de ces couches dans la plaine abyssale.

La Marge Continentale d'Aquitaine. Le bassin de Parentis paraît être une annexe, à l'intérieure du pays, de la Golfe de Gascogne. De là on peut conclure que la formation de la marge a commencé surtout par des mouvements verticaux au début du Crétacé, mouvements qui continuaient jusqu'à la fin du Tertiaire et qui étaient accompagnés de l'affaisement de la région océanique.

Le plateau marginale Des Landes est lié directement au continent. Le canyon de Cap Breton se trouve à la large du sud d'un bloc subsident Tertiaire. La pente continentale de la Golfe du sud de cette région était produite par des failles qui arrivaient pendant la subsidence de ce bloc.

La plaine abyssale. Une auge marginale linéaire, d'une épaisseur de 4000 à 6000 m, s'étend de la dépression de Cap Ferret jusqu'au nord-ouest de l'Espagne. Son profil est asymétrique. L'épaisseur majeure près de la pente s'explique par l'affaisement, continue pendant la sédimentation, et relative au faillage de la pente pendant le Tertiaire et, peut-être, le Crétacé.

Les couches tertiaires et crétacées (?) de Cantabrie et les montagnes sous-marines se perdent vers le sud au sommet d'une élévation. Ces montagnes sous-marines sont des formations tectoniques produites par des failles Tertiaires (post-Eocène à Paléozoique supérieur).

Au nord de Cantabrie il y a une grande région avec moins de sédiments sur une surface accidentée et rugeuse. Dans cette zone centrale, une écorce océanique typique pourrait exister. Le long de la pente Armoricaine se trouve un autre bassin marginale linéaire avec, probablement, des structures perçantes. On compare ces résultats avec des données magnétiques et gravimétriques. On conclut que le Mésozoïque Supérieur doit exister largement dans la Golfe de Gascogne et peut-être des séries Mésozoïques complètes avec, possiblement, des structures diapiriques connexes au sel Triassique.

La formation de la marge par le faillage verticale a commencé entre le Jurassique et le Crétacé et continuait jusqu'à la fin du Tertiaire-Pliocène, avec l'affaisement postérieur de la région de la plaine abyssale et la formation de la pente continentale. On discute ces résultats vis-à-vis des différentes hypothèses proposées.

ZUSAMMENFASSUNG

Man bespricht geophysische und geologische Daten, diese Gegend betreffend. Die wichtigsten Erfindungen der Flexotir seismischen Reflexionsüberbliche, von der Comité d'Etudes Marines, Gruppierung Institut Français du Pétrole, ELF-ERAP, CFP-TOTAL, SNPA-Aquitaine, seit 1967 gemacht, werden dargestellt.

Armorikanischer Kontinentalrand. De Kontinentalschelf scheint eine breite Monokline mit Ausgehen nur von Tertiär, zu sein.

Westlich aber erscheinen unten ältere Horizonte. Man kann etliche Horizonte vom Rand des Schelfes und des Hanges bis an der abyssischen Ebene folgen. Die gegenwärtige Stellung von einem Flachmeer-Albiankalkstein mit einem Kern auf dem Hang versehen beweist eine Senkung von ungefähr 3000 m in der Kreide- und Tertiärzeit. Weiter geschah die Abschiebung von ungefähr 1500 m an Ende des Tertiärs. Diese vertikalen Bewegungen können die gegenwärtigen Tiefe dieser Lagen in der abyssischen Ebene erklären.

Der Kontinentalrand von Aquitaine. Es scheint, dass der Parentis-Becken ein Inlandanhang vom Golf von Biscaya ist. Hiervon kann man schliessen, dass die Bildung des Randes hauptsächlich mit vertikalen Bewegungen am Anfang der Kreidezeit begann und dauerten bis zum Ende des Tertiärs. Diese war mit der Senkung der ozeanischen Gegend begleitet. Der Randplateau von Les Landes ist mit dem Kontinent direkt verbunden. Die Schlucht von Cap Breton befindet sich zum südlichen Rand vom Tertiärsenkungsbloch Verwerfungen, die während der Senkung dieses Blockes geschahen, verursachten den Biscayakontinentalabhang zum Süden dieser Gegend.

Kontinentalrand von Asturien. Eine dicke auf dem Schelf und auf dem Asturienrandplateau gefundene Mesozöische Serie zeigt, dass die pyrenäische Gegend in den Norden von Asturien verlängert. Der Kontinentalhang wurde während des Tertiärs gebildet oder reaktiviert.

Die abyssische Ebene. Ein dicker (4000-6000 m) Linearrandtrog befindet sich von der Senkung von Cap Ferret bis zum nordwestlichen Gebiet von Spanien. Er ist unsymmetrisch in Querschnitt. Man erklärt, dass die grössere Dicke in der Nähe des Hanges folgenderweise verursacht war. Es gab ununterbrochenes Sinken während der Sedimentation, das mit dem Verwerfen des Hanges während der Tertiär- und vielleicht Kreide-zeit verbunden ist.

Die Tertiär- und Kreide- (?) Schichten von Kantabria und die gaskonischen unterseeischen Berge spitzen gegen den Süden auf der Höhe einer Erhebung aus. Diese unterseeischen Berge sind tektonische Gebilde, die von Tertiärverwerfungen verursacht sind (post-Eozän bis zum Spät-Paläozän).

Nördlich von Kantabria befindet sich ein grosses Gebiet mit wenigen Sedimenten, die auf einer ungleichen, gebrochenen Oberfläche liegen. In diesem Zentralgebiet befindet sich vielleicht typische ozeanische Kruste. Des armorikanischen Hanges entlang befindet sich noch ein Linearrandbecken mit wahrscheinlich durchstossenden Gebilden. Man vergleicht diese Resultate mit magnetischen und gravimetrischen Daten. Man schliesst, dass Ober-Mesozöikum in breiten Gebieten vom Golf Biscays sich befinden muss. Vielleicht befinden sich dort ganze mesozöische Serien mit möglich diapirischen Gebilden, die mit Triassalz verbunden sind.

Die Formation des Randes durch vertikales Verwerfen begann zwischen der Jurasischen und Kreide-zeit und dauerte bis zum Tertiär-Pliozän mit folgendem Sinken der abyssischen Ebengegend und der Bildung eines Linearrandtrogs am Fussende des Kontinentalhanges. Man bespricht diese Resultate mit Bezug auf dem verschiedenen Hypothesen, die vorgeschlagen sind.

The continental margin in the Bay of Biscay

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INTRODUCTION

The continental margin in the Bay of Biscay is of particular interest on account of the great geological diversity found in the adjacent continental areas. These areas include stable provinces of young platform (that became an entity after the Hercynian movements) as well as mobile zones during the Mesozoic (Parentis and North Pyrenean basins, the Pyrenees and the folded Basque Arc). This diversity of geological structure enables the continental margin to be divided into different units which will be dealt with successively. At the same time, extensive data from geophysical prospecting are now available and make the Bay of Biscay one of the most closely studied continental margins. Various publications will be cited for each geographical zone considered.

Fig. 1 shows the Flexotir (Cassand and others, 1968; Grau, 1968-1969; Lavergne, 1970) seismic reflection profiles (made by the I. F. P. ship 'Florence') made since 1967 by the Institut Français du Pétrole, the Comité d'Etudes Marines (composed of the Institut Français du Pétrole, ELF-ERAP, CFP-TOTAL and Société Nationale des Pétroles d'Aquitaine) as part of a general research project on the geological structure of continental margins. These profiles are the basis for the results presented here. On this figure are also drawn the Flexotir profiles made recently by the Centre National d'Exploitation des Océans (Noratlante and Nestlante 1 cruises) and by the University of Montpellier (Gestlante III cruise).

MORPHOLOGY OF THE MARGIN AND MAJOR FEATURES OF CONTINENTAL GEOLOGY

The basic data for a study of the margin are provided by the bathymetric surveys made by English (Laughton, 1965) and French oceanographers. We have used a systematic bathymetric survey made by Berthois and Brenot between 1937 and 1964 (Berthois and Brenot, 1962, 1964, 1966a, b) and one recently completed in the south-east of the Bay of Biscay by the Hydrographic Service of the French Navy (SHM maps 6571-6572) (Fig. 1).

Berthois, Brenot and others (1965, 1968) made a morphological and tectonic analysis of these surveys, thus demonstrating the link existing between the morphology of the slope and the geological structure of the inland area.

The following sectors can be distinguished:

1. A unit corresponding to the point where the sedimentary basins in the English Channel are cut off by the slope. This sector will not be discussed in this article.

2. An Armorican unit between 46° 40' and 44° 30'N latitude. The width of the continental slope is between 60 and 70 km and the depth varies from 200 m to 4500-4600 m, thus giving a mean slope of 4°. Submarine canyons are numerous in this area, and their strike is not always exactly perpendicular to the edge of the shelf. The most striking feature is the close parallelism existing between the Vendéean and Breton structural trends corresponding to the major Hercynian shearing movements (Graindor, 1962; Cogne, 1966) and the curvature of the edge of the continental shelf. This parallelism leads to the assumption that ancient tectonic directions played a role in the forming of the continental margin in this area.

3. A more complex Aquitaine unit in which the major features are: the Cap Ferret depression whose northern boundary is made up by the Armorican continental slope which abruptly changes direction here to run east-west; the Landes marginal plateau between depths of 1000 and 2000 m; the Cap Breton Canyon running east-west, with the southern flank making up the Biscay continental slope.

This unit can be connected with the mobile Cretaceous-Tertiary zone indicated on land by the Parentis Basin, the North Pyrenean Basin, the Pyrenean ranges and their westward prolongation in the Biscay region. These elements mainly run east-west.

4. The Asturias and Galicia unit with a continental slope having a different morphology from that of the Armorican sector. It is steeper and has a width of only about 25 km. The canyons are few but large. Their strike varies from south-west/north-east to north-west/south-east and obviously reflects continental tectonic directions. Another significant feature is the existence of the Asturian marginal plateau at a depth of 700 to 1000 m. The eastern boundary of this sector is made up by two large canyons running north-south and

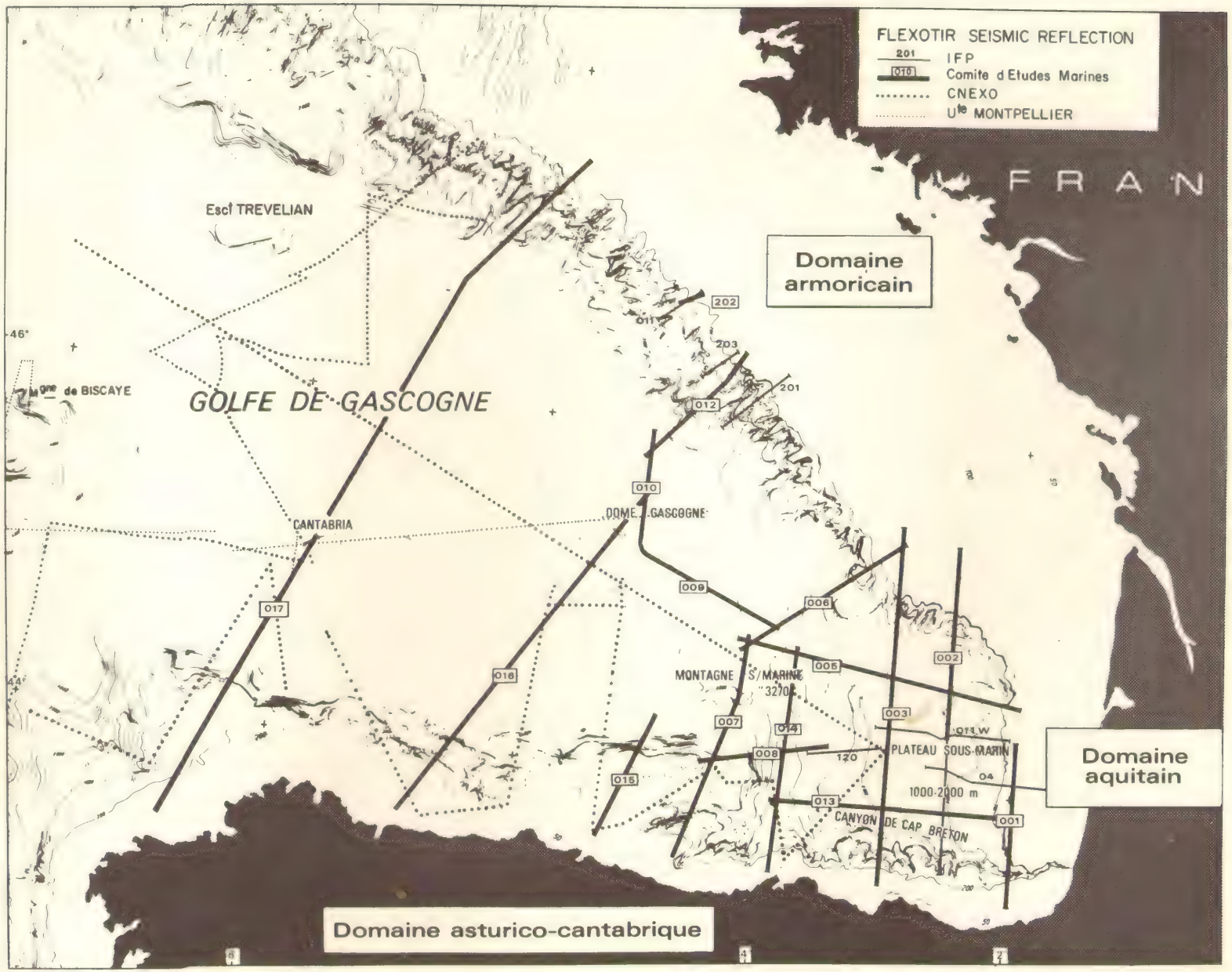

Fig. 1. Bathymetric map of the Bay of Biscay and location of Flexotir seismic reflection surveys.

bounding an intermediate spur. These are the Torrelavega and Santander canyons, with the latter being the prolongation of the Cap Breton Canyon. The continental margin in this unit, which extends over nearly 500 km, can be seen to be arched, with a concavity in the north. On the contrary we have seen that the Armorican continental margin has a concavity in the south.

On shore, the Mesozoic Oviedo-Gijon area apart, outcrops are mainly Palaeozoic and older rocks. Hercynian structural axes make a spectacular virgation, from a south-east to north-west trend to a nearly east to west trend in Asturias.

5. The abyssal region begins at a depth of about 4500 m and becomes gradually deeper as it goes west, up to about 4800 m. The absence of a typical continental rise can be seen.

In this plain, various seamounts have been discovered by surveys (Laughton, 1965, Berthois and Brenot, 1966; Vanney, 1967; Jones and Funnel, 1968). These seamounts are: The Gascony Dome (peak at about 3000 m); Seamount '3270 m' at the foot of the Asturian slope; Seamount Cantabria (peak at about 4250 m) and a few satellites farther east; farther west and almost closing off the Bay of Biscay there are the large Biscay seamounts and, in the north-west, the Trevelyan escarpment which is connected to the English Channel unit.

THE ARMORICAN CONTINENTAL MARGIN

This Hercynian platform area, which extends from 47°30' to 45°N latitude is bounded in the north by the Mesozoic sedimentary basins of the English Channel and in the south by the Mesozoic Parentis Basin and its prolongation offshore, that is, the Cap Ferret Depression.

The Continental Shelf

A great many seismic reflection sparker profiles together with lithological samplings were used to prepare a geological map which is being presented at this Symposium by Boillot and others (all references concerning this area are cited there). In addition, nine refraction lines (Fig. 2) were made by the Compagnie d'Exploration Pétrolière in 1963 as well as a refraction profile southwest of Belle Ile by the University of Montpellier (Martin and others, 1968).

Flexotir profiles 201, 202 and 203 (Cholet and others, 1968) and CEM 012 (Damotte and others, 1968-1969) complete our knowledge of the edge of the shelf. These results as a whole are discussed by Vanney (1969).

Fig. 2 sums up the seismic refraction data and the geological structure of the continental shelf. The sedimentary formations as a whole form a monocline with deeper horizons that appear towards the edge of the shelf. The sedimentary overburden is not very thick, varying from 0 to about 1800 m from Belle Ile to the edge of the shelf, assuming the formations with a velocity of more than 5 km/sec to be Palaeozoic or older.

The age of these formations will be described together with the continental slope. However it is interesting to note that the geological map presented by Boillot and others (this symposium) shows that nothing but Tertiary, mainly carbonates formations, outcrops on this transversal from Belle Ile, from the Eocene lying on top of the basement at Belle Ile up to the Plio-Quaternary on the edge of the plateau. The deep horizons with velocities between 3.3 and 5.2 km/sec are thus probably Mesozoic. These horizons can be seen to pinch out in the south-east (around SR 4) and reappear between SR 6 and SR 9 at the northern edge of the Parentis Basin. Horizons with a velocity of 4.7 km/sec appear in the north-west, showing a thickening of the probable Mesozoic in this direction.

The Continental Slope

We know the structure of the continental slope from the seismic reflection profiles (sparker and air gun) published by Hersey and Whittard (1965), Curray and others (1966), Jones (1968) and Stride and others (1969) as well as from the Flexotir seismic reflection profiles published by Cholet and others (1968) and Comité d'Etudes Marines (Damotte and others, 1968, 1969).

The stratigraphic data obtained by cored or dredged samples are sparse in this area making any reconstruction of the geological history rather hypothetical. Lower Miocene is mentioned between 900 and 1250 m by Hersey and others (1965), as is Lower Miocene (Aquitanian) at 2500 m by Andreieff and others (1968) and by Caralp and others (1969). During the Gestlante III cruise with the 'Jean Charcot' in 1969, Miocene (post-Burdigalian) was found between 3600 and 3800 m. Also three core samples between 2160 and 2375 m, located near the bottom of the I.F.P. 201 profile (Cholet and others, 1968), proved the existence of a gravelly limestone outcrop with algae and orbitolines (Orbitolina aff. discoidea and aff. lenticularis) from the Albian.

Flexotir profiles 201, 202 and 203 (Cholet and others, 1968) revealed the existence of two main lithological groups on the edge of the shelf and at the top of the slope. The

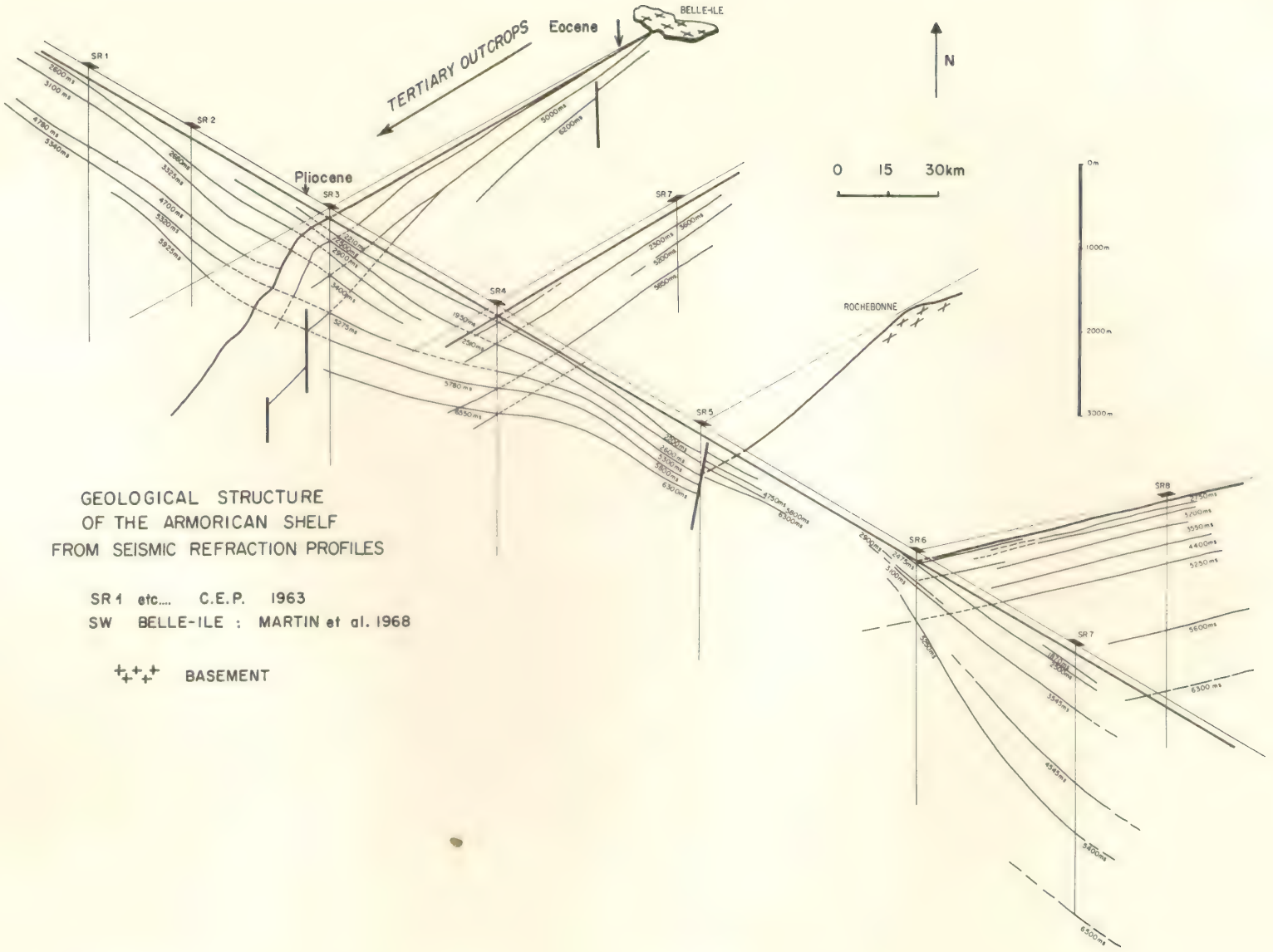

Fig. 2. Geological structure of the Armorican continental shelf from seismic refraction data.

upper sediment group had the structure of a talus prograding offshore, with a velocity varying from 1600 to 2000 m/sec, and forming a wedge from 300 to 600 m thick, with slumping structures taking place on the slope.

The lower group is made up of highly reflecting horizons that are continuous and uniformly thicken offshore, from 800 to 1000 m. These horizons are often eroded by the upper group, and their base is probably an unconformable surface. As shown by the velocities investigations made by Damotte and Hémon (1969, p. 1046, fig. 13), this group in itself is heterogeneous. The upper part has a velocity of about 2500 m/sec, while the lower part has successive velocities of 3000, 2500 then 4000 m/sec. The surface separating these two parts, usually indicated by a high degree of reflection and diffraction, probably corresponds to a geological discontinuity. The total thicknesses found (1600 to 1800 m) are comparable to those in the refraction profile made by Martin and others (1968). The lower group defined here thus probably includes the 2.5-km/sec and 3.3-km/sec horizons revealed by these authors (Fig. 2). The extrapolation of the stratigraphic data from the shelf has already resulted in attributing the 3.3-km/sec horizon to the Mesozoic. This hypothesis now appears to be confirmed by the Lower Cretaceous neritic core samples taken from an outcrop at the bottom of this lower group at the end of profile IFP 201.

Consequently, the stratigraphic succession from top to bottom is probably as follows:

A Plio-Quaternary talus, perhaps already containing some Miocene, deposited by progradation;

A Tertiary group, perhaps including some Upper Cretaceous (Vanney, 1969), increasing in thickness offshore;

Mesozoic horizons which also thicken going offshore, probably containing some Lower Cretaceous.

Profile CEM 012 (Damotte and others, 1968-1969) completes these data. It is of particular significance in that it shows possible correlation between some of the horizons on the edge of the shelf and the abyssal plain (Fig. 3). These horizons are as follows:

a. From the edge of the shelf to a depth of 3000 m. These horizons have been described above and can be summarised thus: Marker 1 – an upper group forming a talus affected by slumping phenomena on the slope, from the Miocene (?) to the Plio-Quaternary; Marker 2 – a lower group becoming thicker offshore, eroded at the top, assumed to be from the Tertiary; Marker 3-4 – an unconformable group underneath 2, from the Mesozoic. Recent vertical faults affect the entire series. Their throw is sometimes as much as 400 m (Cholet and others, 1968, profile 201).

b. From 3000 to 4000 m in depth, the slope increases and the horizons outcrop. A great many diffractions are seen which make it difficult to pick reflectors. The slope break is interpreted as a large faulted zone. This part of the profile can be compared with profile 9 by Stride and others (1969). However, these authors attribute an older age to the different sedimentary groups. The data now available do not appear to confirm this interpretation.

c. Farther offshore, an intermediate block at depths between 4000 and 4500 m, at the beginning of the abyssal plain, shows good reflections that enable a correlation to be proposed for horizons 1 and 2 on either side of the large faulted zone. With this assumption, this major fault would have a throw of about 1500 m. On the other hand, the correlation with horizons 3 and 4 is hypothetical. Faults can be seen in this block which affect only the lower horizons. A flexure accompanied by large faults even further lowers horizons 1 to 4 which are unconformably covered by recent horizontal turbidites from the abyssal plain. It can be seen that horizon 1 does not appear to extend into the abyssal plain and that no fault affects the recent turbidites. Horizons 2 and 4 defined on this intermediate block can be followed underneath the abyssal plain all the way to the Gascony Dome (Fig. 3 profile CEM 010). They will be discussed again in the section 'Abyssal Plain'. Horizons 1 and 2 are probably mainly Tertiary, while 3 and 4 are Mesozoic.

To conclude, the stratigraphic data available and the structural data provided by seismic reflection enable an initial outline to be made of how the continental margin was formed in this area.

It is possible that the Pliocene to Mesozoic horizons on the edge of the continental shelf can be followed all the way to the abyssal plain. The oldest horizons get thicker as they run offshore.

Vertical movements played a major role. The continental slope is the hinge line between the subsident oceanic area and the stable Armorican platform. It is the site of tectonic faults and probably also of important facies changes that become all the more marked as the subsidence of the oceanic area increases. This hinge line was probably originally controlled by an ancient Hercynian fault.

The sinking of the oceanic area is demonstrated in particular by the present position, at a depth of more than 3000 m, of the Lower

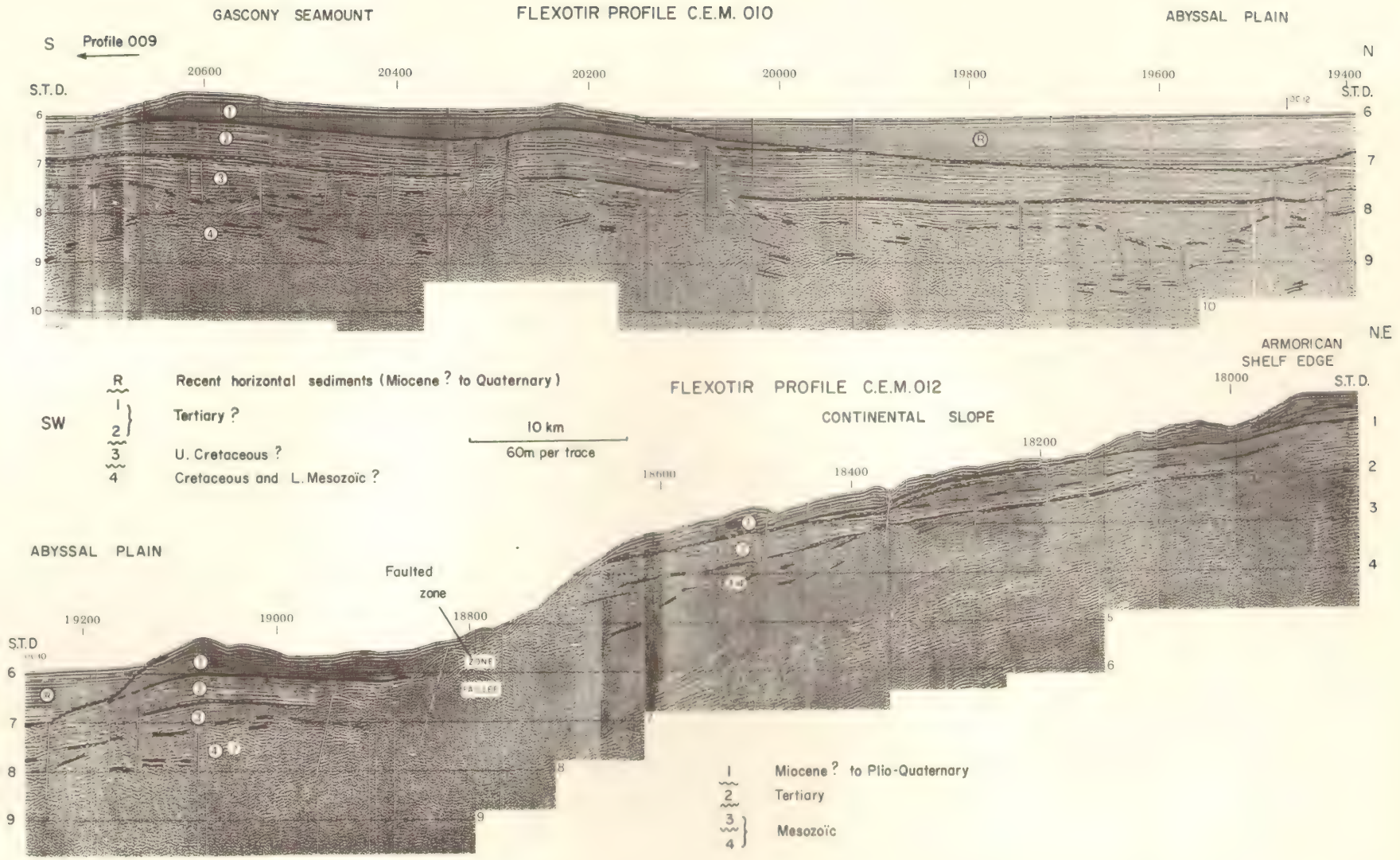

Fig. 3. Geological structure of the Armorican continental slope (profile CEM 012) and its link with the Gascony seamount (profile CEM 010).

Cretaceous with neritic facies; this is quite similar with the conclusion of Heezen and Sheridan (1966) for the Blake Plateau. This tilting movement probably began at least in the Cretaceous, as shown by the unconformity of the Mesozoic underneath the Tertiary and its disappearance eastwards on the Armorican shelf. The Tertiary thickens offshore in the same direction. In addition, extremely large recent vertical tectonic faults (throw up to 1500 m) affect the entire sedimentary series, thus indicating that they are probably from the end of the Tertiary or that they have been seriously reactivated at this epoch. It can be seen that the accumulation of these downward vertical displacements during the geological history of this area can account for the present position of these horizons in the abyssal plain.

The morphology of the slope is not regular, and it is probable that this slope was formed by a succession of 'piano-key' blocks which subsequently governed the location of the submarine canyons.

In agreement with Hersey and Whittard (1965) and contrary to the opinion expressed by Curray and others (1966), we find that in this Armorican region the construction of the margin by sediment progradation is nothing but a very minor phenomenon affecting only the most recent Tertiary sediments.

THE AQUITAINE CONTINENTAL MARGIN

The Cap Ferret Depression and the Western Prolongation of the Parentis Basin

This region is of particular importance for understanding the geological history of the continental margin in the Bay of Biscay, because it is exceptional in providing evidence of a direct link between the oceanic area and a 'continental' sedimentary basin opening up seaward (Figs. 1-14). The geological history of this basin is particularly well known through petroleum exploration work onshore and offshore. This provides an exceptional opportunity for interpreting the geology of the continental margin.

Different sections running north-south on land (Fig. 4A), on the continental shelf (Fig. 4B) and crossing the Cap Ferret Depression (Fig. 5) show us the link between land and deep-waters area.

Fig. 4A is a diagrammatic cross section of the Parentis Basin. A Tertiary and Secondary series more than 5000 m thick can be seen there. Its northern boundary with the Armorican Platform is tied in with the major faults that started to occur at the beginning of the Cretaceous. On the platform the Upper Cretaceous may be seen lying on top of much lower eroded Jurassic layers. Its southern boundary is linked to an arch

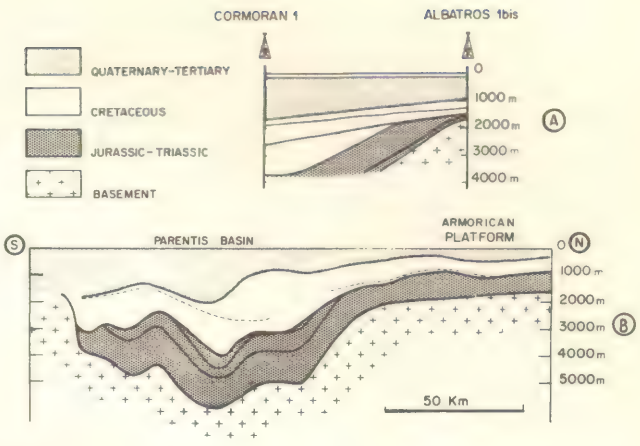

Fig. 4. A. Diagrammatic cross-section of the Parentis basin.
B. Diagrammatic cross-section between two offshore wells on the Aquitaine continental shelf.

separating it from the North Pyrenean Basin (or Lacq Basin) with a great many diapiric structures linked to the Triassic. Cholet and others (1968) have suggested the hypothesis that the continental slope running roughly east-west to the south of the Armorican Platform may correspond to the prolongation of the faulted zone forming the northern boundary of the Parentis Basin.

Two wells on the continental shelf (Fig. 4B) provided an actual demonstration that this boundary continues farther westward. At the same time, the north-south profiles CEM 002 and CEM 003 (Fig. 5) across the Cap Ferret Depression show that the continental slope is effectively the prolongation of the northern edge of the basin all the way to the oceanic area. Therefore, the conclusion can already be drawn that this continental slope has been forming at least since the Lower Cretaceous.

Furthermore, these profiles show that the Cap Ferret Depression was given its present shape by the movement of vertical faults at the end of the Tertiary. The sediments of proven Tertiary age, both on the north flank of the depression (10 core samples taken during the Gestlante III cruise revealed the existence of Middle Miocene: Serravallian and Tortonian, CEM unpublished report) and on the south flank (by continuity with offshore boreholes), have been affected by faults which show increasing throw toward the west although there are none on the shelf. In its present shape the Cape Ferret Depression is thus a sort of graben formed (or with considerable recurrence of movement) at the end of the Tertiary in connection with the sinking of the abyssal plain, as shown by the type of faults bounding it (Fig. 14). A north-south dissymmetry also exists, with the southern part of the depression being more sunken.

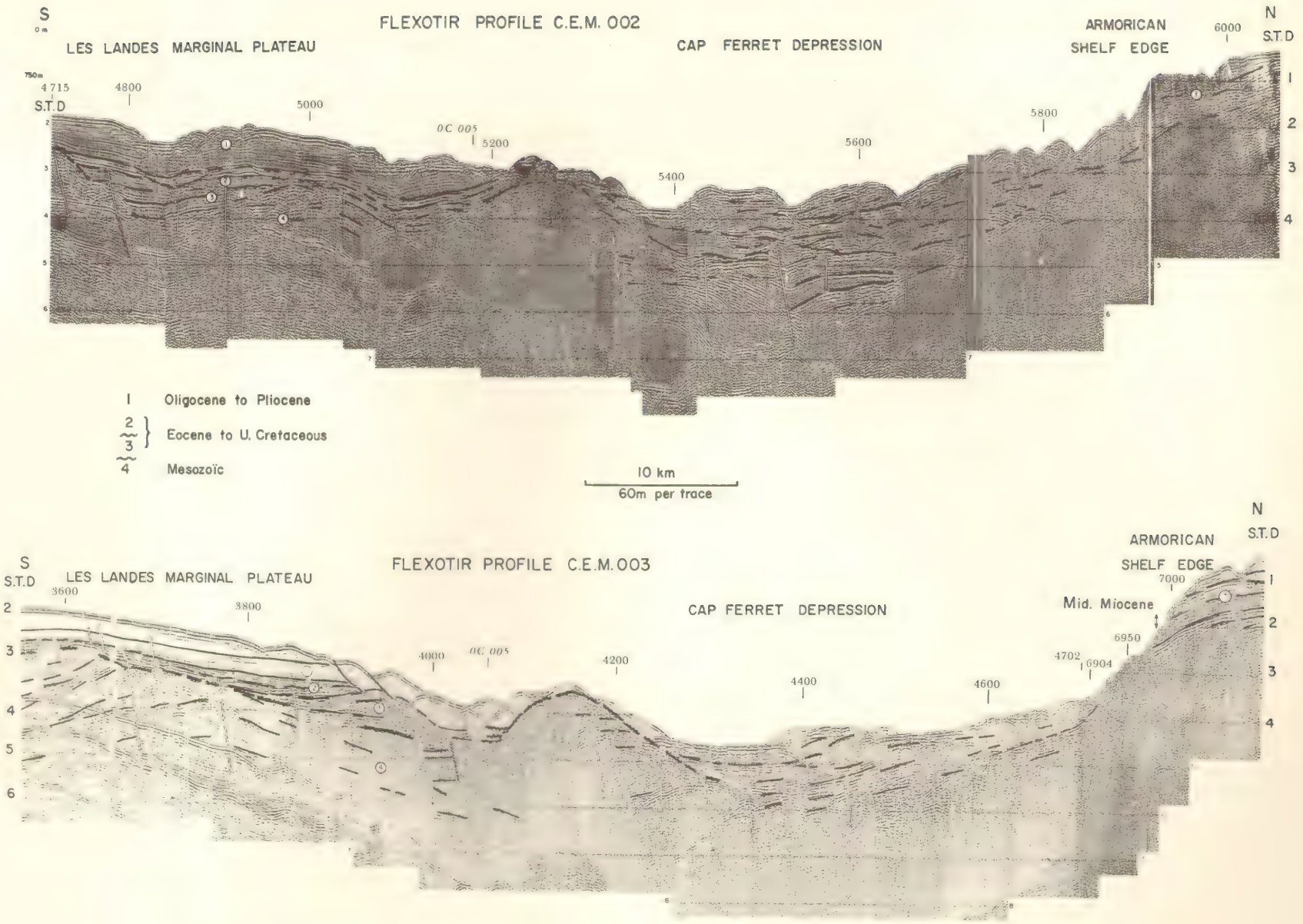

Fig. 5. Seismic reflection profiles across the Cap Ferret Depression (profiles CEM 003 is more westwards than profile CEM 002).

In the southern part of these profiles, the region of the Landes marginal plateau is shown, with deep-lying Jurassic and Cretaceous horizons covered unconformably by Upper Cretaceous to probably Eocene formations and then by a Tertiary (mainly Oligo-Miocene) overburden.

To conclude, it appears that the formation of the Parentis Basin and of the Cap Ferret Depression is closely linked to that of the Bay of Biscay itself. In the continental area, it is the counterpart of the creation of the deepest part of the Bay. This close link between the two areas makes it possible to conclude that the structural formation of the continental margin began with vertical movements, especially at the beginning of the Cretaceous and continuing until the end of the Tertiary, with formation of the abyssal area by sinking. The tensional movements responsible for these vertical faults were thus considerable, as were the compressional movements revealed by the folding of the pre-Tertiary formations.

It can be seen also that the geological history of this part of the margin appears to be the same as that outlined for the Armorican margin although there is less data for the Armorican region.

The Landes Marginal Plateau

This is located west of the Landes continental shelf at a mean depth of 1000 to 2000 m and forms a square, the sides of which measure about 100 km each (Fig. 1). Its northern boundary is formed by the Cap Ferret Depression, its southern boundary by the Cap Breton Canyon, and its western boundary by the Santander Canyon (a prolongation of the Cap Breton Canyon) and the abyssal plain.

Its main features will be summed up here, because much data based on Flexotir seismic reflection profiles have already been published (Cholet and others, 1968; Damotte and others, 1968-1969; CEM unpublished reports).

This marginal plateau is geologically linked directly to the Aquitaine region.

It is covered by an Oligo-Miocene to Recent overburden reaching a thickness of 1000 to 2000 m (Fig. 6 profile CEM 003). This marine clayey-sandstone overburden was deposited by progradation in successive advances of a sedimentary talus. It is a deposit made up of erosion products from the Pyrenees. Towards the east on land, this layer changes to continental facies. It is the equivalent, although deposited in a marine environment opening up towards the west, of the powerful and mainly continental detritic Oligo-Miocene series found south of the Pyrenees. The boundary of the continental shelf corresponds to the present position of this sedimentary talus. Its detailed structure appears clearly on profile 10 by Stride and others (1969).

Mesozoic series are present underneath the Tertiary. They are affected by structures that are probably diapiric, as in Aquitaine (Fig. 6, profile CEM 003). These piercing structures were particularly active during the Tertiary.

Towards the south the Tertiary thickens steadily near the Cap Breton Canyon, and the horizons outcrop on the north flank of this canyon (Fig. 6) (see also profile 11 of Stride and others, 1969). Towards the west, the plateau ends in different ways in the northern and southern halves. In the south, the Tertiary deposits end normally by sinking gradually towards the Santander Canyon.

In the north the sinking of the shelf is linked to a flexure of tectonic origin. One of these flexures lying south-east to north-west stands out clearly in the topography (Figs. 1-14); absent in the middle of the plateau, the throw of this flexure increases towards the abyssal plain until it reaches a total of over 1500 m (cf. profile 120 of Cholet and others, 1968). This folding occurred during the Tertiary and hence has the same characteristics as the recent faults bounding the Cap Ferret Depression.

Still farther north, profile CEM 005 (Fig. 6) will be discussed in the section concerning the abyssal plain. A very large fault linked to a sedimentary trough crossing the entire length of the Spanish margin prevents a direct correlation from being made between the horizons of the Landes marginal plateau and those of the abyssal plain.

The Cap Breton Canyon and the Biscay Continental Slope

The major structural features in this region have been described by Cholet and others (1968) and by CEM studies (Damotte and others, 1968-1969) as determined from Flexotir seismic reflection profiles. In addition, sparker profiles were made at the beginning of the Cap Breton Canyon by Bouysse and others (1968) together with a profile by Stride and others (profile 11, 1969).

Schoeffler (1965a) noted that the Gouf de Cap Breton appeared in the prolongation of a Cretaceous subsident zone north of the Pyrenees. He also showed in particular that the use of isopach maps could reveal the existence since the Oligocene of a long narrow gulf following the axis of the present canyon.

At the same time, the bathymetric map (Fig. 1) shows that the north and south flanks

belong to two different areas. The northern flank corresponds to the Aquitaine continental shelf and then to the Landes marginal plateau, while the southern flank corresponds to the east-west Biscay continental slope with numerous small lateral canyons.

Seismic reflection reveals the following features (Fig. 6, profile CEM 003). From the marginal plateau in the north, the Tertiary sediments thicken steadily to the south to a total of over 2000 m on the edge of the canyon. The upper horizons of these Tertiary sediments outcrop on the northern flank of the canyon, while the lower horizons continue normally underneath, as shown in particular by the 2-3 reflectors corresponding to the bottom of the Tertiary (Eocene to Upper Cretaceous ?). On profile CEM 003 (Fig. 6), the bottom of the Tertiary in the axis of the canyon is thus probably at a depth of nearly 5000 m. These same reflectors shown as 2-3 can also be seen to get deeper, not only from north to south, but from east to west as well (from 3 sec DT to 5.5 sec DT).

The southern flank, on the other hand, is quite different. Although the reflections are not very clear there because of the topography, it can be seen that the layers rise up sharply and are faulted and flexured. The intensity of this phenomenon increases towards the west.

It can also be shown that the Biscay continental slope is linked to a zone of faults or flexures whose throw increases towards the west. This zone was formed between a subsident block during the Tertiary in the north and a more stable Spanish area in the south. The Cap Breton Canyon was probably beginning to be formed in the Oligocene (Schoeffler, 1965a) in this sinking zone. Some authors (Bouysse and others, 1968) feel that the morphology in the eastern part near the continental shelf was caused by fluvial action. It is probable that the Cap Breton Canyon taken as a whole corresponds also to a non-deposition zone whose positioning was governed by structural factors. Therefore, during the deposition of the terrigenous sediments in the north and along the continental slope being formed, there was a constant flow of turbidity currents from the continent to the abyssal plain.

It is worth noting that the east-west fault corresponding to the continental slope cuts across the structural axes associated with the Pyrenean movements occurring in Spain, that is, the folded Basque Arc and the Biscay synclinorium (Mangin and Rat, 1962; Ciry and others, 1967) (Fig. 14). These two structural axes appear to have affected the morphology of the continental slope. However, the Bilbao anticlinorium cannot be followed on the shelf. This clearly shows that the slope is of recent origin and that it cannot be part of a prolongation of a north Pyrenean fault, despite the fact that its positioning is perhaps linked to an ancient structural strike.

This east-west fault appears to extend even farther westward in the Cantabrian and Asturian area, although smaller in size. At the Santander meridian a different structural province begins. It is known on land and is reflected in the submarine morphology. The transition zone between the two provinces is complex. On land the Palaeozoic massif of Asturias is known to plunge deeper towards the east by way of several anticlinal branches lying west-east in the north and north-west to south-east in the south. These anticlinal branches, overthrusted to the south, are linked to movements occurring between the end of the Eocene and the Oligocene. The sinking of this massif in the east is largely due to north-south transversal faults. One of these faults, called the Rio Miera flexure which can be followed to Santander, is especially large (Fig. 14). It continues offshore along the Santander Canyon which prolongs the Cap Breton Canyon towards the north. This then explains why the canyon changes directions. The Tertiary subsident block we have shown is thus bounded in the south and west by tectonic faults.

THE CONTINENTAL MARGIN OF ASTURIAS AND GALICIA

Whereas the outcrops on land are mainly Palaeozoic or older formations except for the Mesozoic outcrops in the Oviedo-Gijon region and the Cantabrian Mesozoic formation in the far east, the numerous sparker profiles and samples made on the continental shelf by the Universities of Rennes and Rouen (Dupeuble and Lamboy, 1969; Boillot and others, 1970) demonstrated the systematic presence of Mesozoic formations, as assumed by Hernandez-Pacheco (1963, 1966). These authors feel in particular that the Upper Cretaceous may be present underneath a flysch facies. From the structural standpoint, an important unconformity exists between the folded Mesozoic and the Tertiary (Fig. 7), the Tertiary series forming elongated basins running west-north-west to east-south-east.

The large canyons cutting into the continental margin are linked to faults visible on land. In particular the Aviles Canyon, running north-west to south-east, can probably be linked to an important structural line which can be followed in the Asturian massif and perhaps on beyond to the Celtiberic ranges.

Flexotir profile CEM 007, supported by cores made during the Gestlante III cruise,

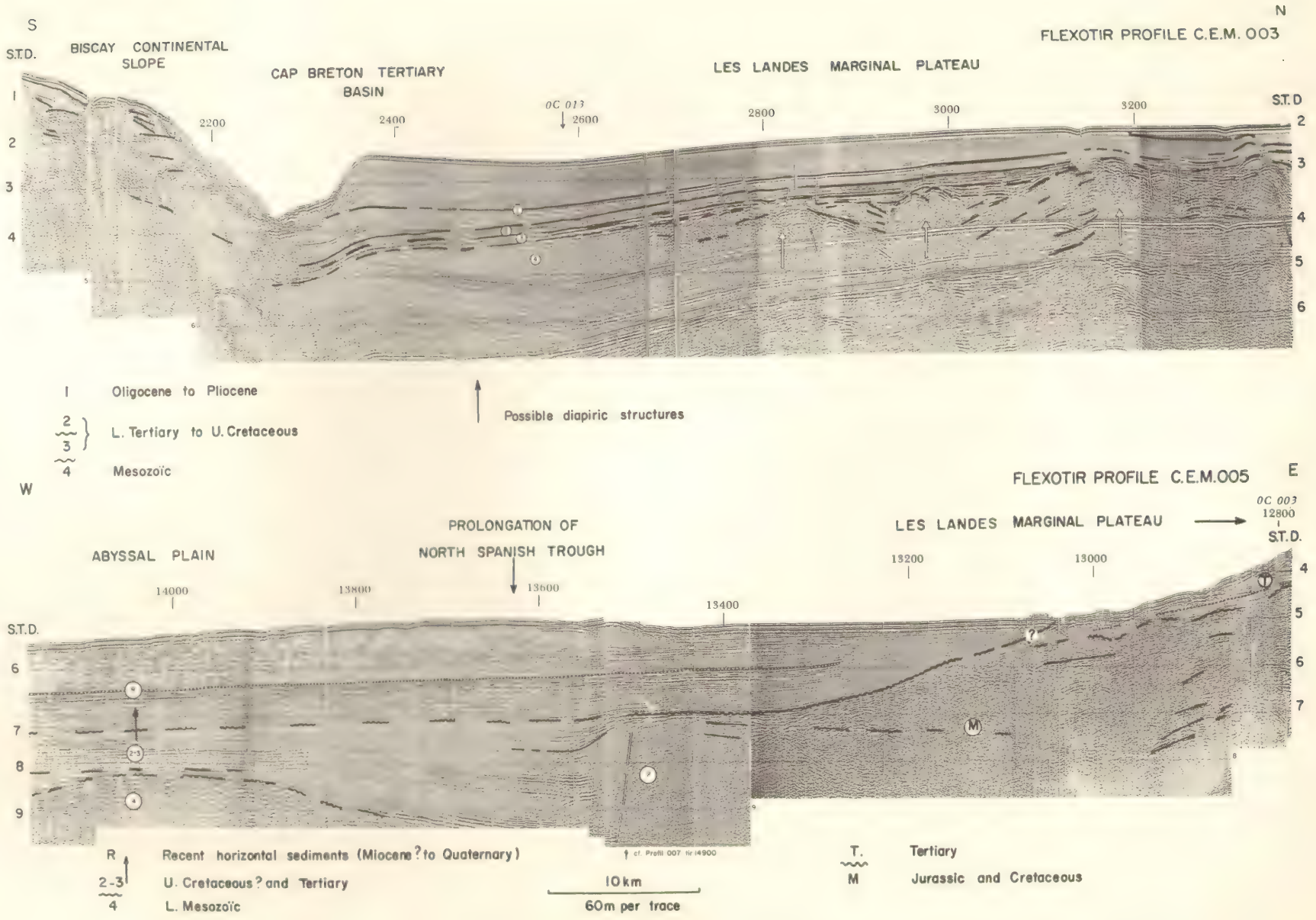

Fig. 6. Profile CEM 003: Geological structure from the Landes marginal plateau to Biscay shelf edge through Cap Breton canyon. Profile CEM 005: Geological relationship between the Landes marginal plateau and prolongation of the north Spanish trough in the abyssal plain

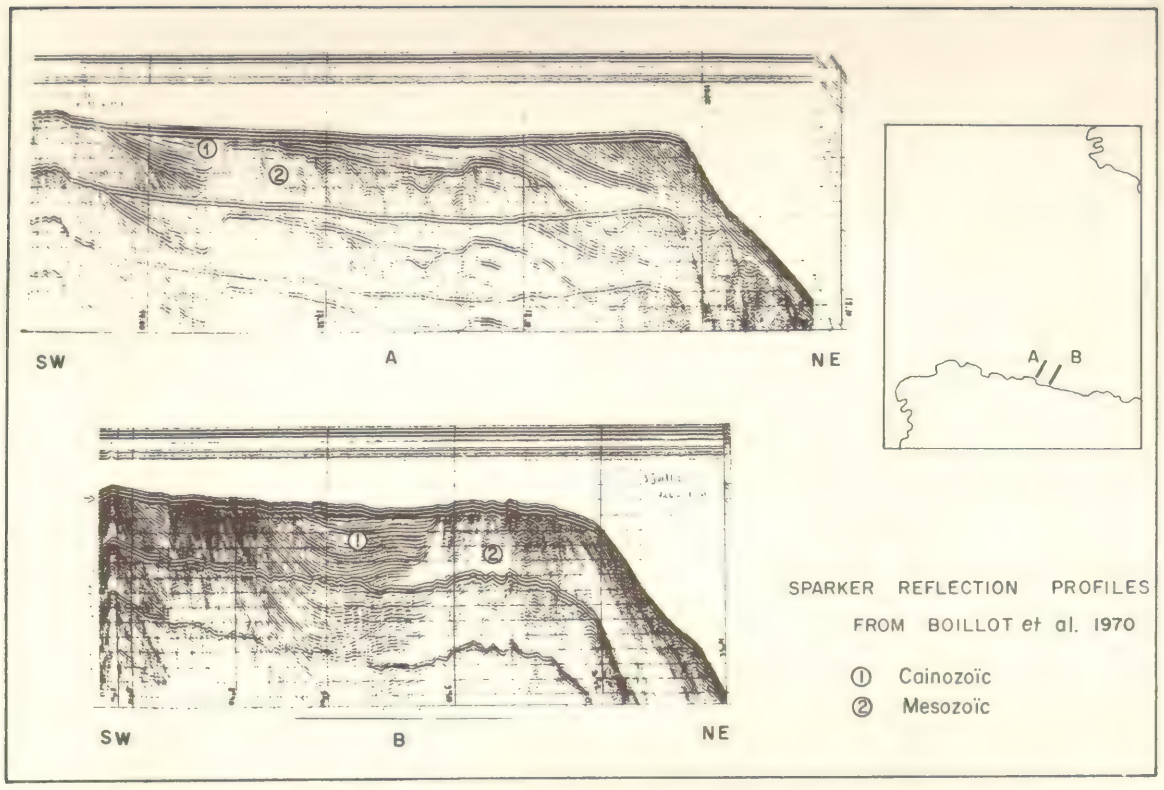

Fig. 7. Geological structure of the Asturian continental shelf from sparker reflection profiles.

provides a cross section of the entire continental margin all the way to the abyssal plain, crossing over the Asturian marginal plateau (Fig. 8). The following main features should be noted.

1. The first slope separating the Asturian continental shelf is linked to a series of faults of Tertiary age.

2. At the foot of this slope there is a Tertiary basin with a canyon running along its axis. On the flanks of this canyon we found Upper Miocene (Tortonian) and Middle Oligocene (Stampian). Underneath this Tertiary, unconformable horizons appear and rise steadily towards the north and then outcrop on the northern half of the Asturian marginal plateau. On the edge of this marginal plateau we cored limestone, dolomite, lignite rocks, some dated from the Lower Eocene to Upper Senonian. Therefore, the unconformity is intra-Eocene and may be linked to the Pyrenean movements. In the middle of the marginal plateau a horst can be also seen that was active during the deposition of the Tertiary sediments. It was caused by vertical intra-Tertiary movements (probably Oligocene).

3. Whereas along Galicia the continental slope is steep, it is sufficiently large and level here for good seismic results. It can be seen that the slope was caused by a series of faults, probably inclined slightly to the north and lowering the Mesozoic and Tertiary formations by blocks. At the top of the slope in particular, one of these faults causes the Mesozoic to have an outcrop nearly 1000 m high.

4. At the bottom of the continental slope, beyond a narrow corridor filled with recent horizontal sediments, lies seamount '3270'. It appears to be a tectonic structure affecting sedimentary formations which are apparently connected directly with those of the Asturian slope. There is a cap of Tertiary sediments which are difficult to link up with the recent sediments on the abyssal plain. Consequently this seamount appears as a sunken block from the continental Asturian area.

5. Beyond this seamount, the abyssal plain begins, with 1.5 sec DT of horizontal sediments lying on top of a blind zone, except for a few reflections apparently connected with a thick series of nearly 2.5 sec DT which continues on northward. Here we can see the same features already found at the foot of the Landes marginal plateau on profile CEM 005 and which exist along the entire length of the Spanish continental slope. This piedmont trough will be described later in the section 'Abyssal Plain'.

Towards the east, seismic reflection shows that the Mesozoic horizons of the Asturian marginal plateau steadily becomes deeper,

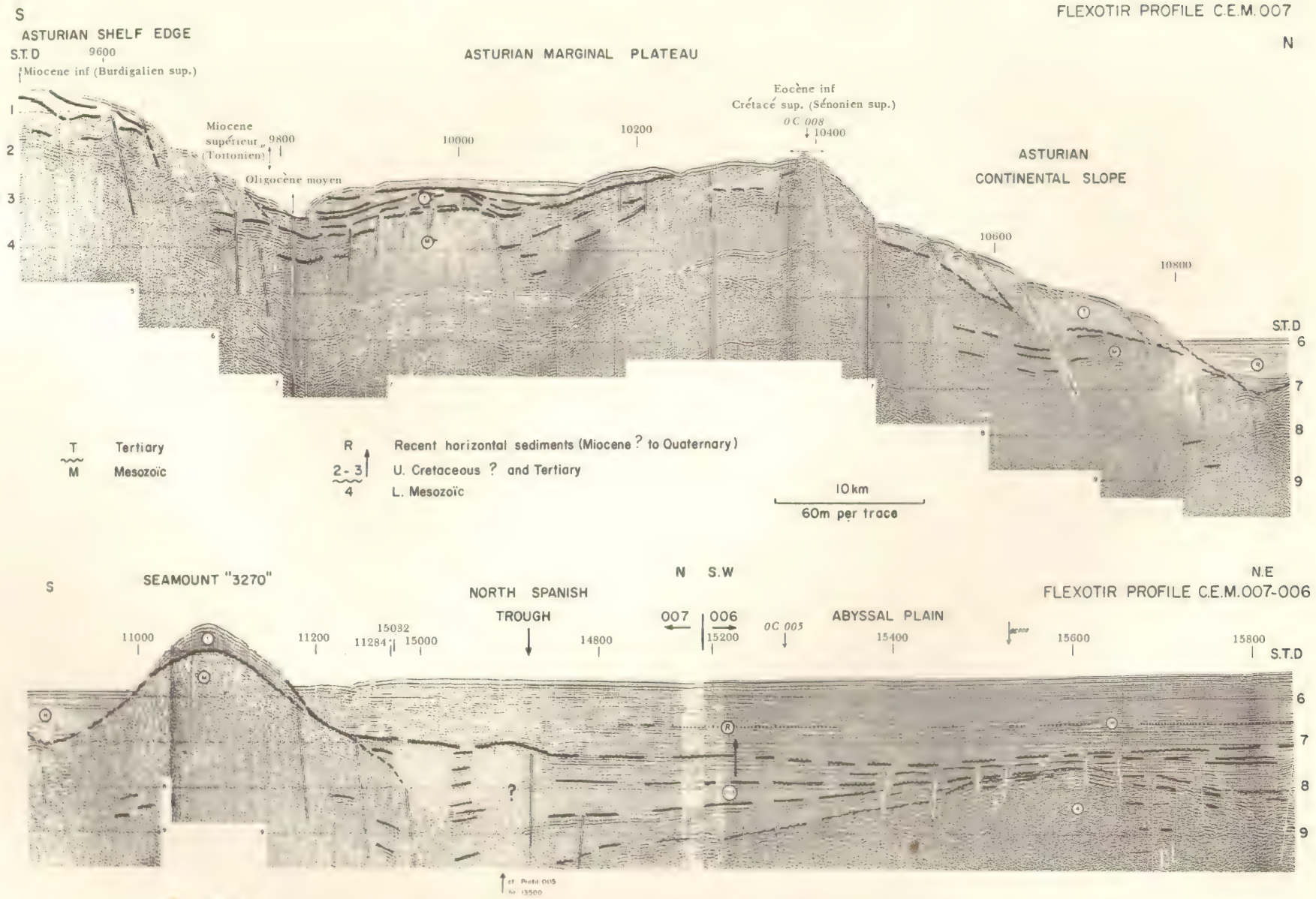

Fig. 8. Geological structure of the Asturian continental margin from the shelf edge to the abyssal plain through the Asturian marginal plateau and the north Spanish trough. (profiles CEM 007-006).

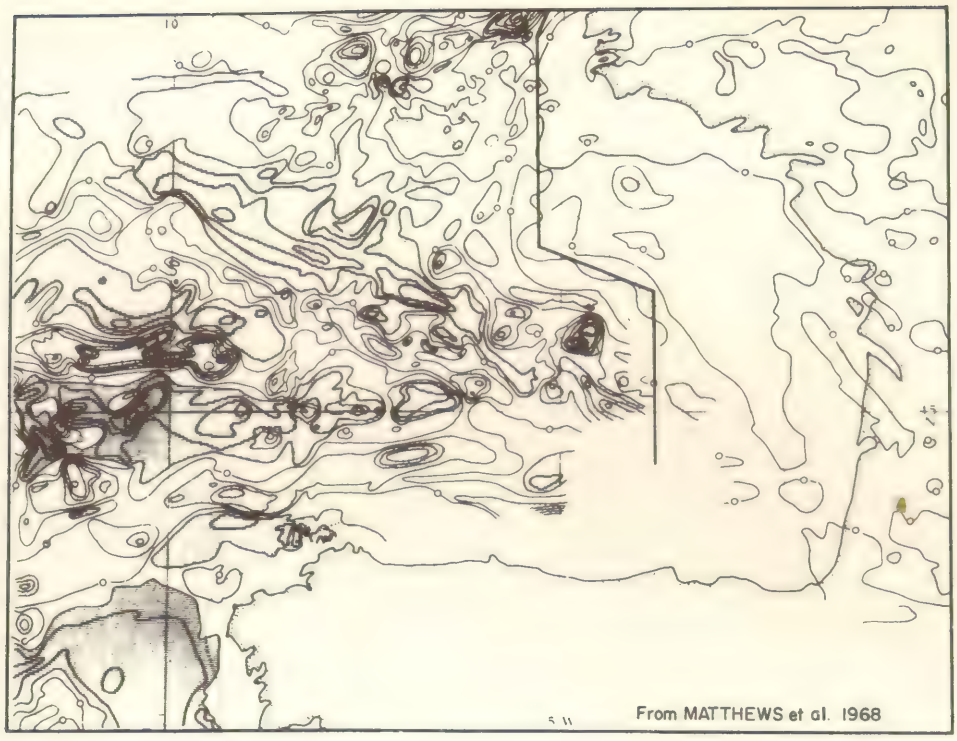

Fig. 9. Magnetic anomalies of the Bay of Biscay (contoured at 50 gamma intervals, dotted areas greater than +50, horizontally ruled areas less than −50 gamma.

Published by permission of North-Holland Publishing Company. From Matthews, D.H. and Williams, C.A. (1968). Linear magnetic anomalies in the Bay of Biscay : a qualitative interpretation. Earth Planet. Sci. Lett., Vol. 4, No. 4, pp. 315-320.

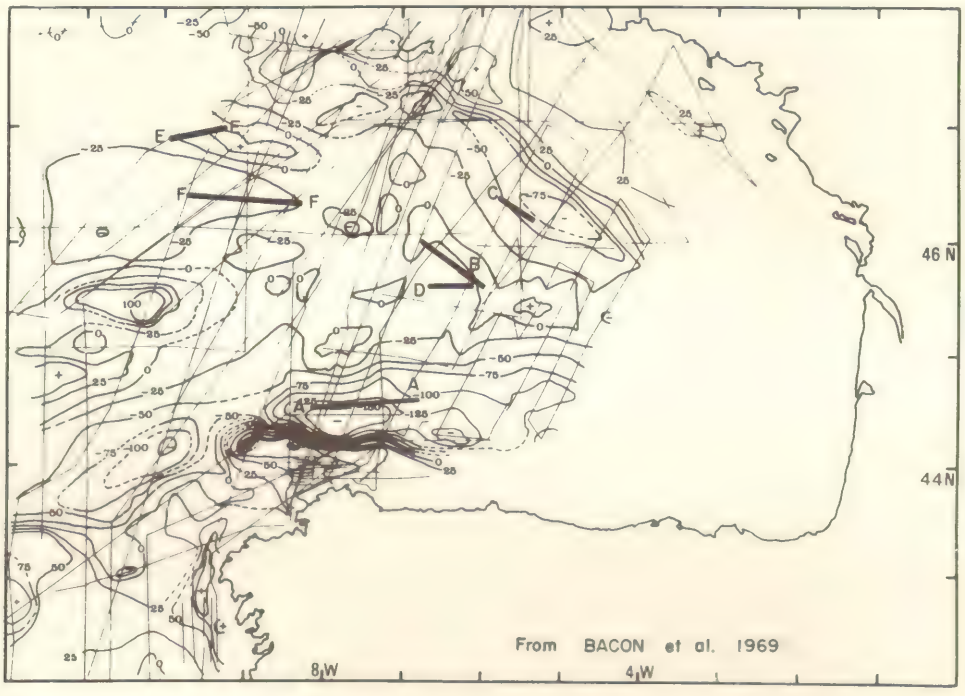

Fig. 10. Free air gravity anomalies (contoured at 25 mgal intervals) and location of seismic refraction profiles (A, B, C, D, E, F).

Published by permission of North-Holland Publishing Company. From Bacon, M., Gray, F. and Matthews, D.H. (1969). Crustal structure studies in the Bay of Biscay. Earth Planet. Sci. Lett., Vol. 6, No.5, pp. 377-385.

almost following the topographic slope towards Torrelavega Canyon. Beyond this, a north-south fault, i.e., a prolongation of the Rio Miera flexure (see preceding section), lowers the horizons even further to the level of the Santander Canyon.

The main geological features now known on the continental margin north-west of Spain can thus be summed up as follows.

Thick Tertiary and Secondary series exist on the continental shelf, the Asturian marginal plateau and the slope. These series are probably the prolongation of those from the Cantabrian province of Biscay. The second feature is a large unconformity from the bottom of the Tertiary on top of folded Mesozoic, probably in connection with the Pyrenean movements. The third feature is the vertical intra-Tertiary movements. Fourthly, the continental slope, which runs for nearly 500 km from 10° W longitude all the way to the north-south break bounding the Santander Canyon (Fig. 14) with a curvature northward, which was partly caused by vertical movements of Tertiary age. These faults probably obliterate more ancient movements about which no data are now available; they are related with the deepening of a sedimentary trough which exists all along the Spanish continental margin.

THE BISCAY ABYSSAL PLAIN

In addition to the magnetic, gravimetric and seismic refraction data to be described below, and the sparker profiles by Jones (1966) and Jones and Ewing (1969), we now have an extensive Flexotir seismic reflection survey (see Fig. 1). The excellent penetration depths attained provide new information on the geological structure of this area. The limited data presented here come solely from the profiles shot by the Comité d'Etudes Marines.

Magnetic Data

Matthews and Williams (1968) published a map of magnetic anomalies (Fig. 9) compiled from towed magnetometric measurements of the total field.

Large anomalies occur here, running north-west to south-east in the upper half of the Bay and east-west in the lower half; while the anomalies in the English Channel run north-east to south-west and end at the edge of the continental shelf.

Matthews and Williams have suggested that this arrangement of anomalies may be linked to basaltic intrusions during a presumed opening in the Bay of Biscay by the rotation of Spain. However, these authors note that a process similar to the one proposed for seafloor spreading is uncertain here, because there is no evidence of a mid-oceanic ridge in the Bay. Furthermore, this hypothesis would also have to include assumption that the injection zone rotated as well. Elsewhere, Le Pichon (1969) said that such a process would have involved the existence of transform faults along small rotation circles centred near 44°N - 4°W.

No matter what hypothesis is accepted with regard to their mode of formation, it is quite probable that these anomalies are linked to the formation of the continental margins and to the process of formation of the deep oceanic structures in the Bay of Biscay.

An aeromagnetic survey of the entire Bay was made by Compagnie Générale de Géophysique for the Centre National de la Recherche Scientifique (Mr. Leborgne) and is scheduled to be published soon. This will provide useful additional data.

Gravimetric and Seismic-Refraction Data

Bacon, Gray and Matthews (1969) published a map of free-air anomalies (Fig. 10) as well as the results of seismic-refraction profiles (Fig. 10 profiles A, B, C and D).

The gravity data revealed the existence of a large negative anomaly at the foot of the Spanish continental slope and another smaller anomaly at the foot of the Armorican continental slope. South of and parallel to the latter anomaly, they suggest the possible existence of a positive anomaly possibly linked to the one revealed on the Trevelyan escarpment (Fig. 10 and 11A).

Seismic-refraction profiles A, B, C and D were made for the purpose of studying the nature of the two marked negative anomalies and of the central quiet zone of the Bay. Using the seismic and gravity data and then compiling a model, the following main features were brought out (Fig. 11): the existence of a thick sedimentary trough containing at least 8000 m of sediments at the foot of the Spanish slope; similarly the existence of an accumulation of sediments at least 5000 m thick at the foot of the Armorican slope; in the middle of the Bay, however the sediments are only 2000 to 4000 m thick depending on whether layers of velocities of 4.6 to 5.3 km/sec are included or not; the structure of the crust in the centre of the Bay is probably of an oceanic type, because there exists a 6.6 km/sec layer and the Moho lies at a probable depth of 11 km. The depth of the Moho probably increases towards the continental margin and reaches 18 km underneath the trough at the foot of the Spanish continental slope. However, these measurements are somewhat uncertain, and if they prove exact this would imply the presence of an abnormal mantle in this region.

A comparison between magnetic and gravimetric data shows that the magnetic anomalies in the abyssal plain are mainly located in the middle part on the Bay where the sediments are thinnest. Only one positive magnetic anomaly was found underneath the sedimentary trough at the foot of the Spanish slope.

It will be possible to make a more accurate interpretation, because the gravimetric profiles made by C.N.E.X.O. with the R.V. 'Jean Charcot' will shortly make it possible to draw up a more complete gravimetric map.

Stratigraphic Data from Seismic Reflection and Sampling

The results of sparker and air-gun profiles have been published by Jones (1966), Jones and Funnel (1968) and Jones and Ewing (1969), but the penetration they obtained in this area is insufficient considering the great thickness of the sediments.

The Flexotir seismic reflection profiles made by the Comité d'Etudes Marines (Fig. 1) enable Damotte and others (1969) to propose a preliminary stratigraphy for the horizons on the abyssal plain and to bring forward new information concerning the geological structure of these horizons. The most recent profiles made by the Comité d'Etudes Marines are still being analysed, and only a few partial results are given here.

RECENT HORIZONTAL SEDIMENTS

CEM Flexotir seismic profiles (Figs. 6, 8, 12, 13) show that the present morphology of the abyssal plain is linked to a thick overburden of horizontal unstructured sediments overlying deeper structured horizons outcropping only occasionally in the form of seamounts (Cantabria, Gascony, '3270 m', etc...).

The thickness of this formation R is uneven, and reflects structures before its deposition. It may attain a maximum thickness of 1800 to 2000 m (2 sec DT), at the outlet of the Cap Ferret Depression. On all the profiles in the south-eastern part of the Bay, formation R can be divided into an upper group about 800 to 1000 m thick (1 sec DT) lying on top of a continuous strong reflector (dotted line). Underneath there are a further 800 to 1000 m of horizontal sediments. When studied in detail it can be seen that the sedimentation of this formation may be complex with discontinuous sedimentary units and what could be described as megaripples (Fig. 6, profile CEM 005), thus conforming well with a turbidite type of deposition. It also appears that the 'upper turbidites', the 'homogeneous zone' and the 'lower turbidites', with a total thickness of 500 m as defined by Jones and Ewing (1969), are found only in the extreme upper part of formation R.

The exact stratigraphic interval represented by this formation has not been determined. Whereas the contact between the upper part of these turbidites and the continental slope sediments appears for the most part to be well defined and unconformable, there is nonetheless a possibility of a link existing between the Tertiary sediments of the slope (see profile CEM 005, Fig. 6, and profile CEM 006-007, Fig. 8) and the lower part of formation R. Profiles in the process of being analysed will probably make this more clear. Consequently, it is possible that the age of this formation ranges from the Miocene, as Jones and Ewing (1969) also suggest, to the Quaternary.

Depending on the location this recent horizontal formation R lies unconformably on top of other horizons (for example, on the Gascony and Cantabria seamount formations) or conformably on others (for example, on the series in the north Spanish trough).

THE STRATIGRAPHIC SECTION OF THE SEAMOUNTS

The Gascony seamount (profile CEM 010, Fig. 3, and profile CEM 009, Fig. 12). The following succession can be seen.

Formation 1. For about 0.6 sec DT (600 m) there is a relatively transparent series that could be similar to the one visible at the foot of the Armorican continental slope. This means that its age may be Upper Tertiary. The formation pinches out underneath Recent sediments on either side of the seamount.

Formation 2. Probably slightly unconformable underneath the preceding formation, its 0.8 sec DT is made up of continuous reflector horizons which can be followed all the way to the Armorican continental slope and be correlated with the Tertiary horizons on the continental slope.

Formation 3. It has good low-frequency reflectors. Its thickness varies from 0 to 1 sec DT, because it fills in a pre-existing topography. It is clearly unconformable underneath formation 2 on the Gascony seamount and almost conformable farther on. Its age is uncertain but is probably Cretaceous (see comparison with Cantabria and description of Armorican continental slope).

Formation 4. It is separated from the preceding one by a major unconformity marked by numerous diffractions. Reflections can be seen there, especially in profile CEM 009. These reflections sketch out large synclines and anticlines which coincide with the irregularities of the unconformity surface. This formation is probably Mesozoic.

The Cantabria seamount (profile CEM 017, Fig. 13C). This location is of particular interest because the vertical movements

have been sufficiently large there to cause deeplying horizons to outcrop so that core samples could be taken from them. In addition, a provisional attempt has been made to correlate them with the horizons on the Gascony seamount by means of seismic reflections.

Jones and Funnel (1969) were the first to define a transparent formation lying on top of the reflectors outcropping on the north flank of the seamount. They obtained core samples showing a conglomerate containing Upper Cretaceous pebbles (probably Maestrichtian) in a Tertiary matrix.

Flexotir seismic profile CEM 017 and the core samples brought back from the Gestlante III cruise in 1969, complete these observations. On Fig. 13C the following can be seen from top to bottom. A relatively transparent formation, with 3 sec DT which is equivalent to the Jones and Funnel (1969) transparent layer and perhaps to Horizon 1 of the Gascony Dome. This formation lies unconformably on top of a formation of about 1 sec DT with strong reflectors in the lower half, perhaps correlating to Horizon 2 of the Gascony Dome. This probably unconformably overlies a formation characterised by a strong low-frequency reflector that may correlate with Horizon 3. Underneath, at about 3.4 sec from the top of Cantabria, there are some steep reflections which might be formation 4.

On the north flank of Cantabria, on the outcrops of formation 2 defined above, three samples of clayey limestone rich in coccoliths and less rich in planktonic foraminifera have been cored at depths of 4660 m, 4797 m and 4788 m. The age is Upper Palaeocene to Lower Eocene.

It can be seen that this finding is in agreement with the assumption of a Tertiary age for Horizon 2 and with the possible presence of the Maestrichtian at some distance underneath. However, some care must nonetheless be taken, because the steep slope of Cantabria corresponds to a faulted zone.

The Stratigraphic Section in the North Spanish Sedimentary Trough

Profile CEM 009 (Fig. 12) shows the progressive pinching out of the 1-2-3 horizons on formation 4 of the Gascony seamount, while the recent sediments R get steadily thicker. Contrary to the opinion expressed by Jones and Ewing (1969), it can be seen that the turbidites on top of formation R are not equivalent to Horizons 1 or 2 of the Gascony Dome.

Profile CEM 007 (Fig. 8) cuts across the above profile and shows that this thinning out of Horizons 2 and 3 is due to the influence during their deposition of an east-west rise of formation 4. In the south, between the bottom of the recent sediments and Substratum 4, the series which is only 0.5 sec DT at the top can be seen to attain nearly 3 sec DT in the trough at the foot of the Asturian slope.

Therefore, it is possible that at least a part of the series filling in the north Spanish trough is equivalent in age to that of the seamounts. The hypothesis of an Upper Cretaceous to Lower Tertiary age can thus be proposed for this lower part. Above, there is a continuous transition with the recent sediments R the age of which being Miocene (?) to Quaternary. An identical and practically total pinching out is found on profiles CEM 017 and CEM 016 farther west (Fig. 13A, B), although the correlation between the horizons of the deep and those of the seamounts is imprecise, a major part of the horizons in the deep clearly appears to be more recent than Horizons 1 to 3.

Therefore, although there is in the axis of this north Spanish deep a continuity of sedimentation between the deeplying horizons of the deep and the recent sediments, there is an unconformity on the seamounts. This reveals an upheaval of the region located to the north of the deep, from Cantabria to the Gascony Dome. The age of this tectonic movement is certainly post-Eocene – Upper Palaeocene (Cantabria).

Another tectonic event occurred during the Tertiary because on all profiles it can be seen that the layers in the southern part of this trough are highly disturbed and then covered unconformably by the unstructured horizontal layers of the formation R (Fig. 13A, B).

The above considerations apparently indicate a Lower Cretaceous and older age for formation 4. Marine Jurassic is known in the Poitou-Vendée region, in Aquitaine up to the Landes marginal plateau, in the provinces of Biscay and Cantabria and in the Oviedo-Gijon Basin. Following the hypothesis on the formation of the Bay, the Jurassic can be present or not in the deeplying parts of the Bay.

Possible piercing structures affecting the deepseated horizons in the abyssal plain in the north Spanish deep (Fig. 8-13A, B) and in the basin at the foot of the Armorican slope (Fig. 13D) may be interpreted as being diapiric structures probably associated with evaporites which, in the regional context, should be probably Triassic. A complete Mesozoic series may thus exist in the deep area of the Bay of Biscay.

The Major Structural Features of the Abyssal Plain

The main structural features thus far revealed by gravimetry, seismic refraction and seismic reflection are shown in Fig. 14

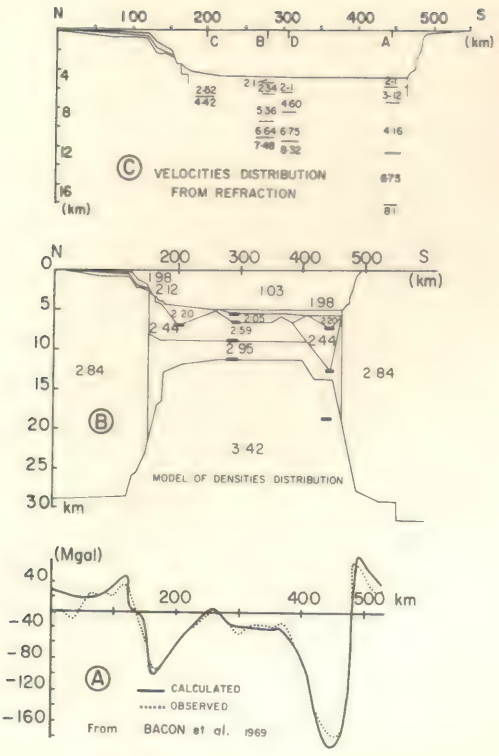

Fig. 11. Refraction data (A) and model (B) combining refraction data and gravity (C).

(A) is published by permission of the North-Holland Publishing Company. From Bacon, M., Gray, F. and Matthews, D.H. (1969)

which is only a preliminary compilation, because a large mass of data is still being analysed. The major structural features are as follows.

THE SEDIMENTARY TROUGH AT THE FOOT OF THE SPANISH CONTINENTAL SLOPE

This deep was revealed by Bacon and others (1969) by means of gravimetry and seismic refraction, and it has been confirmed by Flexotir seismic reflection (profile CEM 016, Fig. 13A, and profile CEM 017, Fig. 13B). Its extent was shown by profiles CEM 007 (Fig. 8) and CEM 005 (Fig. 6) to run eastward all the way to the Cap Ferret depression (Damotte and others 1968-1969). It shows marked dissymmetry with maximum subsidence at the foot of the continental slope. Maximum thicknesses, not counting the recent overlying sediments which may attain 2 sec DT (2000 m), correspond to two-way travel times of 2 to 3 sec. According to seismic refraction (profile A, Fig. 11), the entire sedimentary series is probably about 8000 m thick.

During the Tertiary (after post Eocene-Upper Palaeocene) there was an upheaval of the zone located to the north of this deep, with the formation of seamounts such as the Gascony Dome and Cantabria, both of which are faulted structures. The Biscay mountains perhaps have the same type of structure; Muraour and others (1970) and Frappa and others (1970) interpret them as horsts and mention the existence of Palaeocene outcrops on the flanks just as on Cantabria. In addition to their tectonic structural nature, the seamounts could also have the character of outliers. The series is often more complete at their tops than on their sides. This perhaps shows either erosion on the flanks during the deposition of recent sediments (profile CEM 010, Fig. 3-8 and CEM 009, Fig. 12), or existence of a particular type of deposit on the top while turbidites were accumulated on the deeper adjacent areas.

Independently of this phenomenon, we have demonstrated the existence of a long east-west rise against which the seamount formations pinch out (Fig. 8-13 and 14). This rise formed by formation 4 was active during deposition of these formations.

On all the profiles (CEM 007, Fig. 8, CEM 016, Fig. 13 and CEM 107, Fig. 13B), the sediments in the southern part of the deep can be seen to be highly tectonised by faults and may be piercing structures, perhaps of the diapiric type. It is not impossible that some of these faults are reverse. We have seen that they correspond to a tectonic event of Tertiary age also, but are more recent than the formation of seamounts. An explanation could be that a compression of this sedimentary trough occurred due to a northward movement of the Iberian block. This could be possible because it is known that the main compressional movements in Northern Spain took place from the end of the Oligocene and as late as the Middle Miocene. Thus it can be supposed that the tectonised area of the southern part of the trough is a consequence in the deep area of the Bay, which we know existed already, of the Pyrenean movements taken in a broad sense. The anticlinal structure of seamount '3270', which is included in this tectonised area, might possibly be explained in the same way.

In conclusion, the formation of this linear, east-west trough appears to be linked directly to the formation of the continental margin, with the recurrence of faults on the continental slope increasing steadily with subsidence.

THE SEDIMENTARY BASIN AT THE FOOT OF THE ARMORICAN CONTINENTAL SLOPE

Revealed by Bacon and others (1969), this basin has been confirmed by reflection seismics (profile CEM 017, Fig. 13A). The thicknesses are not as great as in the north Spanish deep. They correspond to two-way travel times of about 3 to 3.5 seconds, including the recent sediments but excluding the deeperlying series. A large proportion

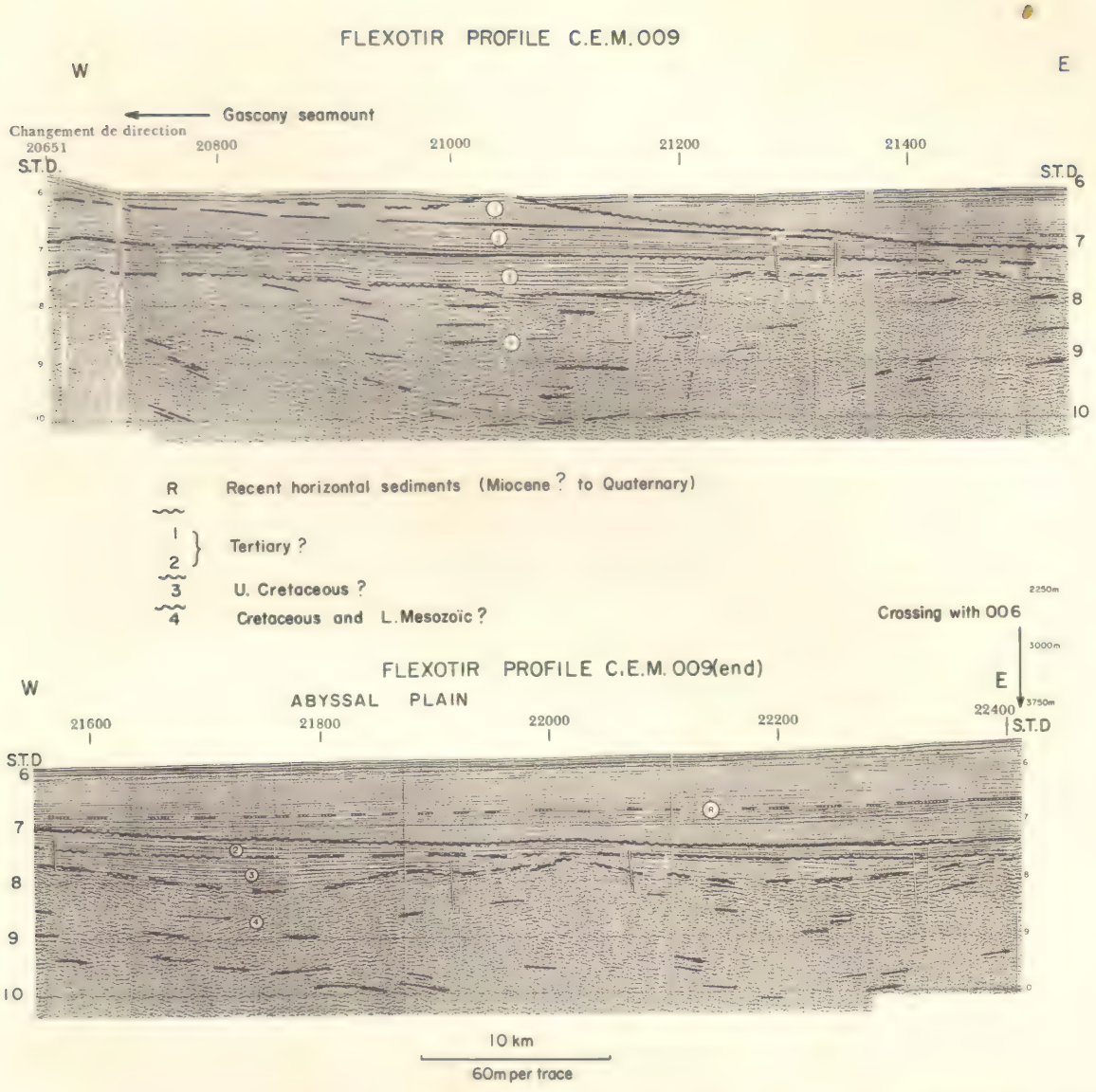

Fig. 12. Stratigraphic section in the abyssal plain south-east of the Gascony seamount (profile CEM 009).

of these sediments appears to be equivalent to formations 2 and 3. Piercing structures, perhaps diapiric, seem to exist. It is probable that the relative importance of the different stratigraphic levels are different in the two marginal basins. Particularly Oligo-Miocene formations could be much more important in the north Spanish trough, in relation to erosion of the growing Pyrenean relief.

Like the north Spanish trough, its formation appears to be linked directly to the formation of the continental margin. However, its cross section is slightly different in that it is less dissymmetric. Whereas in the north Spanish trough, the south flank was associated with the synsedimentary occurrence of a large fault forming the continental slope, here the landward boundary appears to be linked to faulting but also to a flexure of the slope as a whole. This explains the morphological opposition which we have pointed out between these two slopes. Moreover, there is no tectonised part in this marginal basin as in the north Spanish one.

THE CENTRAL PART OF THE BAY

Seismic refraction in this area shows the sediments to be much less thick, that is, about 4000 m, including formations with velocities of 4.6 to 5.3 km/sec. Seismic reflection confirms this finding (Fig. 13C) and, underneath the recent sediments and those that may be attributed to formations 2 and 3 for which the total thickness corresponds to two-way times of 1 to 2 seconds at the most, reveals a highly diffracting surface with a great many faults (top of formation 4?) and a highly irregular topography containing numerous structures that appear to be piercing in some cases. The fact that the sedi-

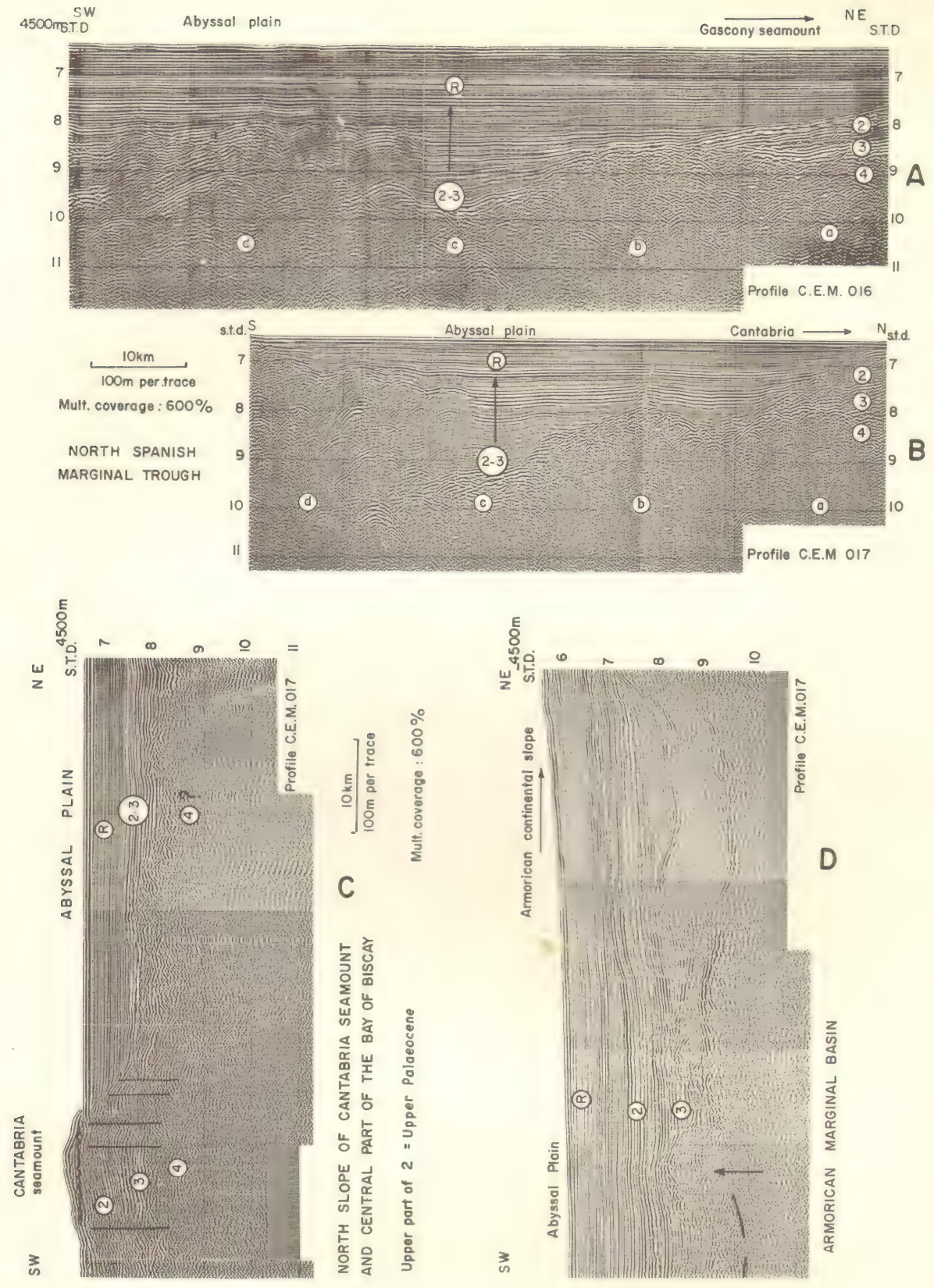

Fig. 13. A and B. North Spanish trough south of Gascony and Cantabria seamounts. From north to south, (a-b) pinching out of the Seamount Series on an east-west ridge of formation 4, (c) undisturbed series of the trough, (d) tectonised part of the trough. C. Cantabria seamount and central part of the Bay of Biscay. D. Armorican marginal basin with possible piercing structures (arrow).

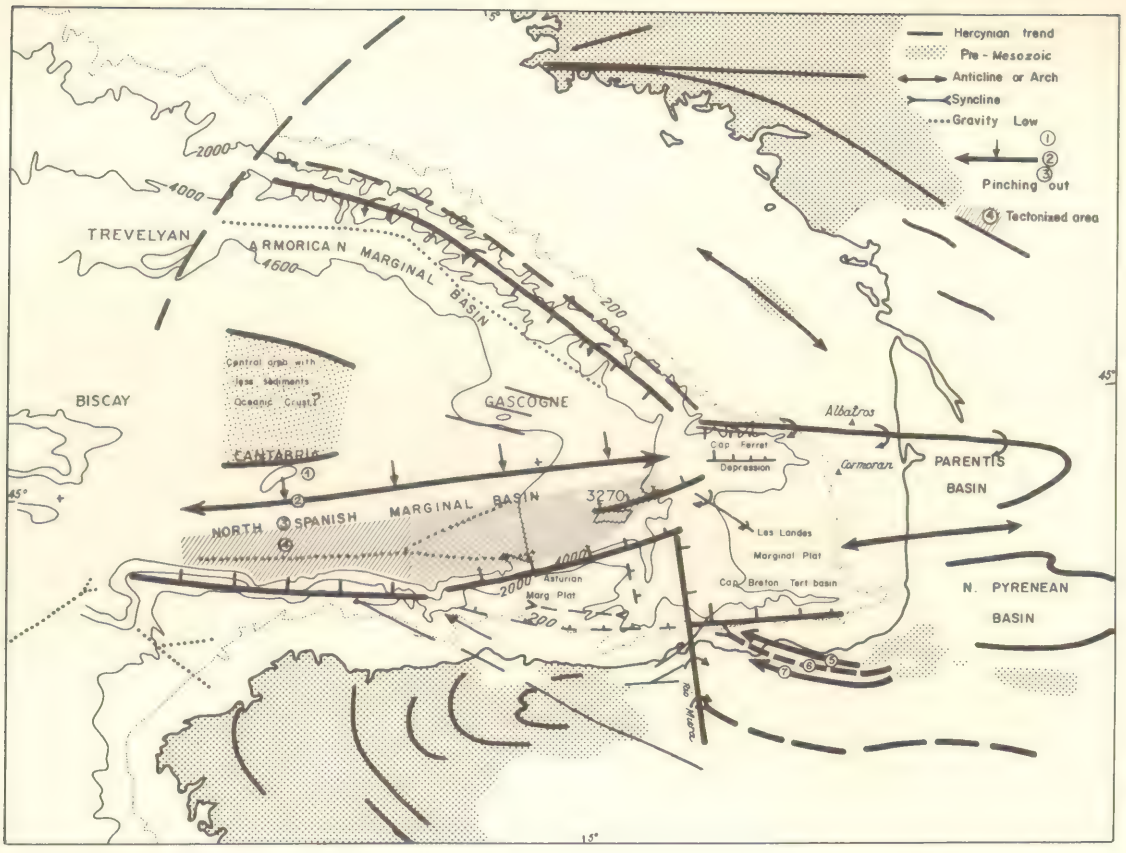

Fig. 14. Main structural features of the Bay of Biscay. 1. Area uplifted during Tertiary with typical seamount series. 2. Ridge of formation 4 on which seamount series pinch out. 3. Undisturbed area of the north Spanish trough. 5. Basque Arc. 6. Biscay synclinorium. 7. Bilbao anticlinorium.

ments are not as thick can be interpreted as being linked to the greater distance from the continental margins, with the sediments tending to accumulate in the form of linear basins at the piedmont of continental slopes. The deeplying structure is probably typically oceanic, and the magnetic anomalies stand out especially sharply (Matthews and Williams, 1968).

To conclude, as on adjacent land areas it can be seen that the structural trends of the southern half of the Bay of Biscay lie east-west and that they take on an Armorican trend in the northern half.

CONCLUSIONS CONCERNING THE FORMATION OF THE CONTINENTAL MARGIN

A great many hypotheses have been put forward. We shall review them before comparing them with new findings and reaching provisional conclusions.

The Different Hypotheses

The most commonly accepted hypothesis is that Spain rotated by about 30° in a counter-clockwise direction around a point located somewhere west of the Pyrenees or of Aquitaine. This would account for the triangular shape of the Bay. Carey (1955) proposed this theory as did Bullard and others (1965). The latter, in order to fit the limit of continents prior to the formation of the Atlantic (along the 500-fathom contour lines), felt that there had to be a junction of the Spanish and Armorican continental slopes. This rotation was what explained the formation of the Pyrenees (Schoeffler, 1965b) and theoretically enabled the Hercynian trends in Galicia and Brittany to be connected.

New arguments in favour of this hypothesis have been provided by comparing the results of palaeomagnetic measurements on both sides of the Pyrenees, especially those made by Girdler (1965, 1968), Van der Voo (1967) Van Dongen (1967), and Watkins and others (1968, 1970). More recently, Van der Voo (1969) has reviewed all this research and has concluded that there was a 35° rotation of Spain between the Triassic and the Lower Cretaceous, with the better measurements

coming from Lower Permian rocks.

In addition, the revealing of magnetic anomalies in a fan-shaped pattern (Matthews and Williams, 1968) together with the oceanic type of structure of the middle part of the Bay of Biscay (Bacon and others, 1969) are important facts that are often brought up in support of this hypothesis.

However, the hypothesis has been contested by Mattauer (1968a, b, 1969) who proposes, as the origin of the oceanic depths in the Bay, a movement of the Iberian block of about 400 km away from Europe along a right-hand transcurrent fault. He points out that the connection between the Hercynian structures in Galicia and Brittany is debatable and, in particular, that it was impossible to reveal any shortening of approximately 300 km in the eastern part of the Pyrenees as would have been necessary if an opening up of the Bay occurred in the Lower Tertiary. The presence of Upper Cretaceous in the Bay as mentioned by Jones and Funnel (1968) makes it even more difficult to accept the supposed link between the formation of the Pyrenees and the opening up of the Bay. For Mattauer, formation of the oceanic area would be then related to the Hercynian period.

CEM seismic reflection data (Damotte and others, 1968-1969) revealed the extent of the vertical movements linked to the continental margin, the subsidence of the abyssal area and the existence of a complete sedimentary series in the south-eastern part of the Bay.

Recent Data and Discussion

The main features revealed by Flexotir seismic reflection are summed up and their consequences discussed below.

The continental slope forms a hinge zone between a continental area and a subsident oceanic area. It is not merely a sort of scar separating two entirely independent regions. Consequently, it has undergone extensive vertical faults and probably also facies changes.

Data now available show that vertical movements have begun at least at the beginning of the Cretaceous and have continued until the end of the Tertiary. In connection with the formation of the continental margin, thick sedimentary piedmont basins were formed at the foot of the Armorican and Spanish continental slopes. Thus these features seem very similar to those described for the north-eastern American continental margin (Drake and Nafe, 1967; Drake, 1969; Drake and Kosminskaya, 1969). As a result, the Cretaceous and Tertiary are probably extremely widespread in the Bay.

The presence of older Mesozoic formations is possible, at least in the vicinity of the margins. The marine Jurassic is found along the entire perimeter of the Bay of Biscay and in Portugal, implying the indisputable presence of an Atlantic Ocean, even of reduced size, in the Jurassic. Thus this Jurassic may be included in formation 4 which is the deepest one revealed. The Triassic may also be present if a diapiric origin is attributed to the piercing structures in the abyssal plain. In the light of the subsidence of the abyssal plain and considering the probable existence of a complete series, it seems that the Bay of Biscay was much shallower at the beginning of the Cretaceous than at present.

The almost proven existence of the Mesozoic in large areas confirms that the formation of the Pyrenees cannot have been linked to an opening-up process of the Bay. The history of the Bay of Biscay most probably included several episodes of tension and compression.

For example, formation 4 appears folded in the south-eastern part of the Bay. But the main compression episode on land was obviously connected with the formation of the Pyrenees s.l. during Tertiary. At that time the abyssal plain must have been mostly formed already, and traces of this northward movement of the Iberian block should exist. A hypothesis is proposed that the tectonised area affecting the southern part of the north Spanish trough may be linked to some compressional movements during the Tertiary. It must be emphasised that there is no such tectonised area on the Armorican marginal basin.

The hypothesis that the Bay opened up would imply that this movement took place entirely before the Mesozoic, because otherwise Mattauer's (1969) objection that it is impossible to find a shortening in the eastern land areas by nearly 300 km of the Mesozoic would still be valid. Moreover oceanic crust area created in such a way may not have attained its present depth until later, by subsidence during the Mesozoic. It is evident that Mattauer's objection should be carefully checked, because in this instance land geological data in favour or against this hypothesis have more weight than oceanic geophysical data, especially magnetics. The latter can be interpreted differently according to the hypotheses that are adopted.

Because of the probable presence of Mesozoic formations around the abyssal plain, Jones and Ewing (1969) have proposed the existence of a south-east to north-west segment of oceanic crust, formed after the Cretaceous by a 12° opening up of the Bay. Considering data now to hand, the proposed theory is no

longer valid for the south-eastern part of the Bay where the structural features lie east-west. The problem of the middle part of the Bay remains, however, with its thinner sediment layer and its oceanic crust. Either the uneven diffracting surface existing there is linked to a volcanic substratum and the hypothesis by Jones and Ewing may be partially correct if sediments upon it are only of Tertiary age, or else this surface corresponds to consolidated sediments (velocity of 4.6 to 5.3 km/sec) and the thinness of the underlying sediments may simply be linked to the great distance from the continental margins, with the greater thickness being limited to the marginal basins. The rotation hypothesis, even when limited, always comes up against the same problem of the shortening that would have to be found in the opposite sector.

To conclude, the formation of the continental margin by an opening up of the Bay comes up against various limitations with regard to the age (probably pre-Mesozoic) and the process (subsidence of an initially shallow oceanic substratum). This can be opposed just as easily by another hypothesis maintaining that the Bay of Biscay is a former continental area that sank during the main period of formation of the northern Atlantic Ocean, but this would imply: a conversion of the continental crust into an oceanic crust according to a hypothetical process, such as has been suggested for other parts of the world (Drake and Kosminskaya, 1969; Drake, 1969) while the magnetic anomalies would then be the result of this process; alternatively a westward shift of the initial line along which the Atlantic Ocean opened up such as is proposed by various authors (Drake and Nafe, 1967). In this case, the triangular shape could be explained by the influence of Hercynian structural trends.

Consequently, the facts presented are interpreted in a hypothetical and provisional manner. It is obvious that any further advance in our knowledge will require: (1) additional stratigraphic data which, it is hoped, will be provided by the JOIDES programme, despite the difficulty due to the presence everywhere of thick sediments; (2) the use of recent advances in seismic data processing, because of errors that may have been made in interpreting seismic sections, especially in steeply dipping areas (Fontanel et Grau, 1969a and b; Fontanel and others, 1970); (3) a synthesis of all available offshore data (CEM and C.N.E.X.O. seismic profiles interpretation still in progress, latest cruise by Cambridge University) and a comparison with land geological data, because we feel that the two areas did not have entirely independent geological histories.

This subject will be dealt with in a symposium on the structural history of the Bay of Biscay which the Comité d'Etudes Marines, C.N.E.X.O. and Cambridge University are proposing to organise.

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Morphology, sediments and Quaternary history of the continental shelf between the Straits of Dover and Cape Finisterre

G. Boillot, P. Bouysse and M. Lamboy

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SUMMARY

A study of the sediments and morphology of the continental shelf off the French coast has brought to light the Quaternary history of this shelf. Regressions have left important formations which are today submerged: fluvial, aeolian and periglacial deposits, valleys or karstic depressions, notches resulting from coastal erosion.

Today, the relict deposits are more or less altered by tidal currents. Present day sediments are rare and rather thin; they comprise pelites in certain areas of the Bay of Biscay and calcareous organogenic sands north and west of Brittany, which sometimes form sand-ripples and hydraulic dunes.

The history of the Spanish continental shelf is not so well-known. Glauconite is abundant.

SOMMARIE

Une étude des sediments et de la morphologie de la plate-forme continentale le long de la côte de la France a éclairci l'histoire quaternaire de cette plate-forme. Des régressions ont laissé de formations importantes qui sont noyées aujourd'hui: des dépôts fluviaux, éoliens, et périglaciaires, des vallées ou des dépressions kartiques, des encoches formées par l'érosion littorale.

Aujourd'hui, les dépôts qui restent sont plus ou moins changés par des courants de la marée. Les sédiments actuels sont rares et assez minces; ils comprennent des pélites en certains endroits du Golfe de Gascogne et des sables calcaires organogéniques au nord et à l'ouest de la Bretagne, qui forment quelquefois des rides de sable et des dunes hydrauliques. L'histoire de la plate-forme continentale espagnole n'est pas si bien connue, la glauconie abonde.

ZUSAMMENFASSUNG

Eine Untersuchung der Sedimente und der Morphologie des Kontinentalschelfes abseits von der französischen Küste hat die Quaternärgeschichte dieses Kontinentalschelfes ans Tagelicht gebracht. Regressionen haben wichtige Formationen gelassen, die heutzutage wasserbedeckt sind. Diese Formationen sind: Fluviatit, Äolische- und Periglazial-Ablagerungen, Täler oder Karstsenkungen, Hohle, die aus Küstenerosion folgen.

Heutzutage haben Gezeitenströmungen die hinterbliebenen Ablagerungen mehr oder weniger abgeändert. Jetzige Sedimente sind rar und etwas dünn: sie bestehen aus Pelit in gewissen Bereichen von der Bucht von Biscay und aus kalkartigen organogenen Sänden zum Norden und Westen von der Bretagne, die kleine Sandwellen und Hydraulik-Dünen manchmal bilden.

Die Geschichte des spanischen Kontinentalschelfes ist nicht so bekannt. Dieser ist reich an Glaukonit.

Morphology, sediments and Quaternary history of the continental shelf between the Straits of Dover and Cape Finisterre

G. BOILLOT, P. BOUYSSE and M. LAMBOY

THE CONTINENTAL SHELF OFF THE FRENCH COAST

Introduction

The English Channel and the Celtic Sea (west of Brittany) are subject to strong tidal currents. Off the French coast they exceed 1.2 m/sec at the surface and may locally reach 2.5 to 5 m/sec during spring tides, allowing only pebbles to exist on rock floors. In the gulfs and bays, for instance in the Bays of Mont-Saint-Michel (Bourcart and Boillot, 1960) and of the Seine (Larsonneur, 1969a), these tidal currents are weaker and allow the deposition landwards of successively finer sediments, from gravels through coarse and fine sand to mud in the estuaries.

In the central part of the western English Channel, south-west and west of Brittany, as far as La Chapelle Bank and beyond, pebbles give way to coarse sands which are to a great extent organogenic and calcareous (Berthois, 1946; Boillot, 1965). In these areas a tight series of sand ripples (Stride, 1963) show that currents continue to be an important factor in sedimentation.

South of Brittany, tidal currents are much reduced and only occasionally affect the sea bottom. In an elongated area extending from the Penmarch Point to the Rochebonne rock outcrop and corresponding approximately to the middle of the South Armorican Massif, lies the Grande Vasière (the great muddy area). The term must not be taken literally: in reality, the lutite content of the sediment rarely exceeds 50 per cent and the mud generally forms only a thin veneer over the sand (Berthois and Le Calvez, 1959).

Along the South Armorican Massif mud also occurs in a series of littoral basins protected from the open sea swell by a succession of islands and rock outcrops which separate them from the shelf itself: Concarneau and Lorient basins (Bouysse and Le Calvez, 1967); Mor Bras and Vilaine basin (Vanney, 1969; Bouysse and others, 1966); and the Bay of Bourgneuf (Gouleau, 1968).

The rest of the continental shelf of the Bay of Biscay is to a great extent covered by medium or fine sands, mostly siliceous (with the exception of the northern half in the vicinity of the outer edge, where they are calcareous) and slightly clayey. From north-west to south-west, up to the Gouf de Cap Breton, the average grain size decreases (Lapierre, 1969a, 1969b). At the same latitude, sands are coarser towards the shelf edge than in its central part. This is due to two phenomena: the absence of a modern supply of detritus beyond the Grande Vasière leads to the persistence of coarse residual sediments enriched in organogenic calcareous sediments and perhaps also by the action of currents due to internal tides in the vicinity of the shelf edge.

Features Inherited from Pre-Pliocene Events

From Cap de la Hague to the Vendée, the Armorican Massif is surrounded by the sea and many morphological features on the continental shelf are inherited from Hercynian structures. Such is the case of the coast line itself and the succession of islands and rock outcrops in southern Brittany which separate the shelf from a series of fore-coastal basins.

It is also probably the case for the shelf edge which, from 48°N to the area off the Gironde, curves parallel to the structures of the South Armorican Massif. It is therefore possible that the continental slope itself is the result of a late revival of Hercynian tectonic deformations (Berthois and others, 1965).

The shelf is formed of Secondary and Tertiary sediments (Fig. 3) where reliefs are rare: Eocene 'chicots' (knolls) south-west of Groix (Pinot, 1966a); Eocene banks on either side of the mouth of the Loire (Vanney, 1969); the Great Barrier, west of the Vendée, which is a small and discontinuous elevation formed by Miocene deposits (Klingebiel and others, 1968).

Here and there, however, banks of unlike constitution appear: these reliefs are older than the surrounding Mesozoic and Caenozoic formations, for example, in the Bay of Biscay, the crystalline rocks outcropping at Rochebonne and Ile d'Yeu.

Nearer the coast, the Hercynian basement emerges from the post-Palaeozoic cover. The boundary between the two geologically different areas is very clearly marked in the submarine morphology: the calcareous region is a plateau terminated sometimes by a structural, some-

Fig. 1. Morphological sketch of the French continental shelf. After Boillot (1964), Bouysse and Horn (1968c), Hinschberger (1963 and 1968), Hommeril (1967 and 1968), Larsonneur (1967 and 1970), Pinot (1966a and 1970), Stride (1968), and Vanney (1968 and 1969).

times by an erosion surface. The Hercynian region has a generally irregular morphology which continues the continental topography under the sea. The very upper limit of the continental shelf lies at the contact of the basement and the post-Hercynian cover. It is 100 m deep west of Brittany, but shallows to the north-east and south-east, rising at least to the level of the Cotentin or Noirmoutier. Further east, the shelf continues through the Paris and Aquitaine Basins.

The general picture is of a westward tilting since the deposition of the Tertiary cover (Eocene and Miocene). The seawards dip of the shelf and consequently the depth of the shelf edge (-180 m), seem to be fairly recent features. This movement caused the submersion of a part of the Armorican plateau. The submerged topography results from the Tertiary reactivation of the Hercynian structures, particularly of faults, and from a differential erosion of rocks (Guilcher, 1948).

Relict Plio-Quaternary Deposits

PEBBLES, COBBLES AND CONTINENTAL OR LITTORAL GRAVELS OF THE ENGLISH CHANNEL

Pebbles and associated siliceous gravels abound in the region south of the western and eastern English Channel, where tidal currents remove sand and clay. They are found also in certain areas west and south-west of the Armorican Massif, mixed with sand as far as the shelf break and even deeper. In the Gulf of St Malo, as in the other regions where Hercynian rocks occur, a close relationship exists between the petrographic nature of the pebbles and that of the bedrock (Hommeril, 1967; Larsonneur, 1969b). Elsewhere, especially in the western English Channel, with the exception of chert, most of the pebbles are of a completely different nature from that of the underlying formation. They are imperfect cobbles or little rounded rock fragments originating in the Armorican Massif, while the shelf is formed of Secondary and Tertiary calcareous rocks.

The morphology of most of the pebbles shows that they have undergone frost weathering. At the shelf edge, they overlie Pliocene deposits (Andreieff and others, 1969). Both these facts suggest a Pleistocene age.

Their deposition has been explained by the melting of drift-ice (Dangeard, 1928). Though this phenomenon may have had some influence at the time of transportation of boulders now scattered over the shelf, it does not explain the great number of pebbles in the English Channel nor their occurrence only in the topographically deepest parts of the Channel. On the other hand, some recemented fragments of split cobbles on the present bed of the Channel testify to in situ frost action (Boillot, 1964).

Clearly most of the cobbles were deposited in the open air during a period of regression. Those situated deepest in the Channel and on the Celtic shelf seem to be of fluvial origin and were split by frost after deposition. Others, probably more recent, were deposited by solifluction under a periglacial climate in the areas near the Armorican Massif.

It is quite remarkable that when the sea returned to its present level it did not generally alter the shape of these pebbles, which have preserved almost unchanged the features of previous periglacial action. But the general rule allows a few exceptions: in the Gulf of St Malo and off Ouessant, well rounded cobbles, derived possibly from old beaches, occur at various depths (Hommeril, 1967; Hinschberger, 1969).

SHELLY CONGLOMERATE IN THE ENGLISH CHANNEL

Some pebbles and gravels in the Channel are cemented by a limestone which may contain a marine fauna very similar to the present day fauna. Several arguments indicate that this fauna lived during a Pleistocene transgression phase. Submarine cementation cannot be observed under present day conditions, and on the contrary, rock samples indicate destruction of the cement. Further, the occurrence of loess in some of the cemented shells shows that the shells were submerged and covered after deposition by a continental formation (Lefort, 1969-70). The shelly conglomerate is therefore not Holocene, and at least part of the pebbles (the cemented ones) predate the last regression. A second deposition of pebbles might, however, have occurred during Würmian times.

COBBLES, GRAVELS AND CONGLOMERATE IN THE BAY OF BISCAY

Cobbles and gravels are also found in many places on the shelf in the Bay of Biscay (Berthois, 1955). Some are of coastal (Pinot, 1968), others of fluvial origin. Finally a shelly conglomerate containing a little characteristic fauna has been observed on several occasions (Vanney, 1969). Formations similar to those of the English Channel can therefore be said to exist in the Bay of Biscay.

SILICEOUS SANDS IN THE BAY OF BISCAY

Reddish brown sands cover the shelly conglomerate and stretch over the entire plateau north of the Bay of Biscay to the Celtic Sea (Berthois and Le Calvez, 1959; Lapierre 1969a, 1969b; Andreieff and others, 1970a). The siliceous fraction is composed of quartz grains with a diameter greater than 200 μ rounded with traces of aeolian action and red

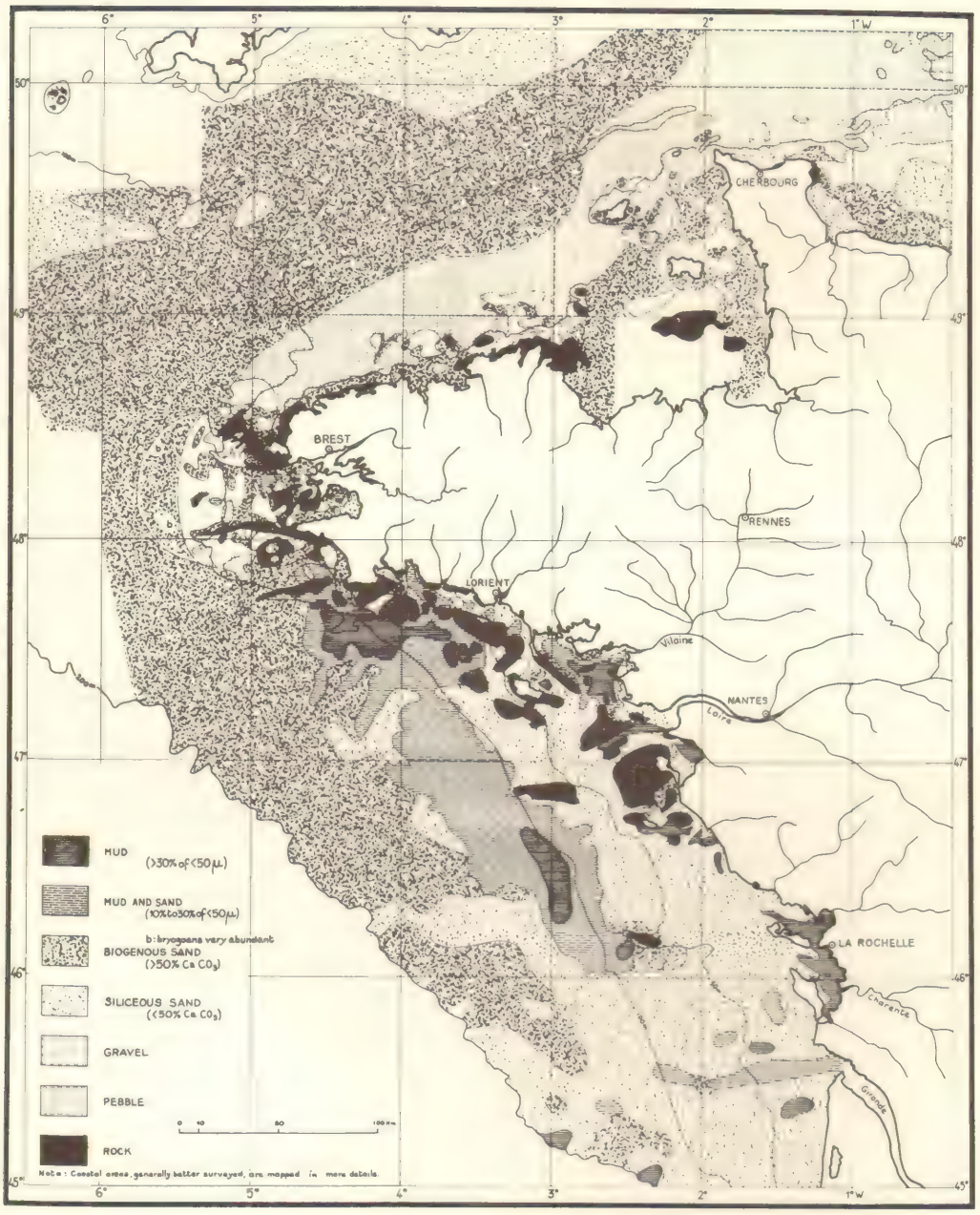

Fig. 2. Schematic distribution of superficial sediments off the western shelf of France. After Andreieff and others (in press), Boillot (1964), Bouysse and Le Calvez (1967), Duplaix and Boillot (1968), Hinschberger (1963, 1968), Hommeril (1967, 1968) Lapierre (1969a), Larsonneur (1967, 1970), Saint-Requier (1969), Vanney (1968, 1969).

or brown colour due to ferric oxide. They lie on the shelf in strips which seem to stretch from the mouth of the Loire and more clearly from the Gironde. Hence the idea that these are old fluvial deposits corresponding to a regression and subsequently more or less altered by either sea or wind action.

This hypothesis is confirmed by the study of heavy minerals contained in the sand. The mineralogical associations characteristic of the Loire and Gironde can be more or less easily followed from the present river mouths to the shelf edge. The reddish brown sands are therefore visible evidence either of the pre-Flandrian or of an earlier regression.

Grey sands in the Bay of Biscay are different: angular, without iron staining, fine grained (average diameter less than 200 μ, usually 100 μ), micaceous, with a particular mineralogical composition. They are doubtless more recent than the reddish brown sands and probably contemporaneous with the successive stages of the Flandrian transgression.

ORGANOGENIC REMAINS AND MICROFAUNA

The fossil or ancient character of the organogenic remains is seldom established. Some absolute age measurements have been attempted on supposed littoral shells extracted at different depths on the shelf. These gave: at -188 m, 9000 years B.P.; at -50 m, 5100 years B.P.; and at -32 m, 3000 years B.P. (Lapierre, 1969a).

Further north, in the Bay of Audierne, a deposit at -34 m yielded shells dated at 15 000 B.P. (Saint-Requier and Guilcher, 1969). Even if this last result is surprising compared to what we know in other cases, nevertheless, shells and organogenic debris of present sediments may be very old. In this deposit the most numerous shells are evidence of the successive stages of the Flandrian transgression, as, for example, the yellowish organogenic remains mixed with the reddish brown quartz sands (Bourcart, 1947). But one must not dismiss the possibility of finding still older shells, freed, for example, from the shelly conglomerate by present day dissolution of the calcareous cement.

The benthic foraminifera mixed with sediments between -170 and -130 m and between -140 and -120 m lived respectively at shallower depths, of some 50 m. Only foraminifera extracted beyond the shelf or from between 0 and -120 m seem to be contemporary, on ecological grounds (Andreieff and others, 1970a).

Relict Plio-Quaternary Morphology

CLOSED DEPRESSIONS

Especially in the English Channel but also to the south of Brittany, closed depressions of two different kinds are found: some are elongated troughs, among which the best known is the Central Trough (Hurd Deep), others, much smaller in length, are funnel-shaped with more or less circular or elliptical edges.

The diameter of the small circular depressions ranges from tens to hundreds of metres. The bottom reaches 100 m beneath the deepest part of the sea bottom, which is some 30 m beneath the contiguous plateau.

In every case, the depressions are cut in calcareous ground and thus may be compared to dolinas eroded in an emerged plateau. Several arguments point to the conclusion that this karst was formed in the Pleistocene during a regression but later than the principal scattering of pebbles.

The Croisic trough lies off the mouth of the Loire and is 10 km long and 1 or 2 km wide (Vanney, 1964). A light seismic reflexion survey (Bouysse and Horn, 1968) showed that it is a section of the submarine valley of the Loire which has not been in-filled by sediments locally.

Many troughs are known in the Western Channel. They are very similar, in spite of their diverse dimensions (Table 1) and those who have studied them have attempted to find a common explanation for their genesis, which is perhaps a mistake.

Four hypotheses have been proposed:

1. The troughs are sections of the Seine Valley where there has been no deposition. The Hurd Deep especially is very similar to a continental valley and even shows nested channels (Boillot, 1963c). It has not been established, however, that the troughs are the continuation on the shelf of valleys filled with sediments; seismic-reflexion surveys have even shown that the Ouessant and Ile Vierge troughs are rocky at their extremities and have no hidden prolongations at their upper and lower extremities (Boillot and Horn, 1966; and unpublished data).

2. The troughs are tectonic deeps (Dangeard, 1928). There is a close parallel between the direction of the Hurd Deep and that of magnetic anomalies (Hill and Vine, 1965). This is not proof that the troughs are rift valleys, but rather indicates that the location is structurally imposed (Barthe and others, 1967).

3. The troughs result from present day excavation by tidal currents (Stride, 1963). There are, however, many places in the Channel where currents, though violent, have not caused the formation of depressions and inversely, troughs occur where currents are not strong. The role of tidal currents mainly prevents the in-filling of depressions. However, hydrodynamic conditions were perhaps different when the sea level was lower than it is today.

Fig. 3. Provisional geological map of the French Atlantic shelf and the Channel. After Andreieff and others (1969, 1970b), Barthe and others (1967), Boillot (1964), Bouysse and Horn (1968c), Hinschberger (1963, 1968), Hommeril (1967, 1968), Horn and others (1969), Larsonneur (1967, 1970), Lefort and Deunff (1970), Stride and others (1969), Vanney (1968, 1969).

Table 1

	L (km)	l (km)	d (m)	p (m)	ROCK FORMATION
Ouessant trough 1. (Barthe and others, 1967; Boillot and Horn, 1966)	16	3.0	100	200	Cretaceous and Tertiary limestones
Ouessant trough 2. (Hinschberger, 1969)	6	0.6	30	130	
Ile Vierge trough (Barthe and others, 1967; Boillot, 1964)	20	1.5	40	130	limit Green Chalk-Lower Cretaceous
Pluteus trough (Boillot, 1963a)	27	2.3	20	90	Chalk and calcareous sandstone
La Hague trough (Boillot, 1963b; Robert, 1969)	15	1.8	50	110	limit chalk-basement
Central trough (Hurd Deep) (Boillot, 1963c; Robert, 1969)	165	6.5	100	175	various Jurassic, Cretaceous and Tertiary formations (Fig. 3)

L - length; l - width; d: difference of level between the plateau and the bottom of the trough; p - maximum depth of the trough; 1. main trough; 2. adjacent trough.

4. The troughs are karstic forms hollowed out during regressions (Boillot, 1963a, 1963b, 1963c, 1964). It is a fact that all the Channel troughs are located in calcareous areas (except for the upper part of the Hurd Deep, excavated in rocks which are locally argillaceous), and it is also a fact that all possible morphological intermediates between the important elongated depressions and the small circular depressions, which are probably of karstic origin, can be observed, but the dimensions of the Hurd Deep are a serious objection to this interpretation.

5. An explanation has been proposed for the Ouessant Trough: that the depression results from old hydro-aeolian erosion, such as certain present day depressions in the Gobi desert (Hinschberger, 1969). This explanation seems to us to be of difficult general application.

It appears from this enumeration that the hypothesis that the Channel troughs resulted from structurally guided sub-aerial erosion (valleys and polje) is the most probable. One must consequently consider a regression of over 200 m (depths of the Ouessant trough) at the time of their formation. The troughs of Ouessant, Pluteus, La Hague and the southern part of the Hurd Deep lie in a vast region covered with Pleistocene pebbles but are nevertheless not in-filled. One must conclude that the troughs were formed after the scattering of the pebbles, that is, during the late Pleistocene.

SUBMERGED OR BURIED RIVER SYSTEMS

To a depth of 120 m, the continental hydrographic network can be traced on the continental shelf (Fig. 1). Valleys are sometimes apparent in the present day topography but are more often in-filled by sediments and require seismic-reflexion work for identification. (Bouysse and Horn 1968, Robert 1969).

How old are these valleys? Morphological data on the adjoining emerged area indicate possible Pontian formation (Vanney, 1969). Their in-filling is probably contemporaneous with the great Quaternary scattering of material (pebbles and reddish brown sands, etc.).

EROSION LEVELS, SUBMERGED BEACHES

There are numerous submerged shores and notches cut in the sedimentary bed rock which is the seawards continuation of the Hercynian basement (Table 2). Although authors explain them generally by prolonged periods when the Flandrian sea was stationary, it is possible that some of these shorelines may be older.

Present Day Phenomena

SEDIMENTARY FORMS

Vast sand-ripples and sand-ribbons lie in groups or singly everywhere in the Channel

Table 2

	a	b	c	d
St Malo gulf (Hommeril, 1967)	-	52-55*	37-45*	28-15*
North of Western Channel (Cooper, 1968)	-	-	40	-
South of Western Channel (Lefort, 1969-70)	-	46-50	28-42	14-25
West of Brittany Celtic Sea (Hinschberger, 1969)	90-100	50-55	-	-
South of Brittany 1. (Pinot,1966b,1968)	89-92	52-55	37	-
South of Brittany 2. (Vanney, 1969)	106 -	-	40-45	18-20

a. b. c. d: superimposed shore lines (m)

1. - off the coast from Penmarch to Belle-Ile

2. - off the coast from Belle-Ile to the mouth of the Loire.

The levels followed by an asterisk are indicated only by pebbles; the others, by erosion notches, with or without submerged littoral deposits.

and south of the Celtic Sea, especially at the Chapelle Bank. Some of the ripples are fixed and were formed under different conditions from those prevailing today (Berthois, 1959; Berthois and others, 1969). Others are produced by the currents which work on the sea floor (Carruthers, 1963).

The asymmetry of the modern ripples may serve to determine the direction of sedimentary transit. Today the sands drift in an east-west direction from the western part of the Channel to the south of the Celtic Sea and perhaps in the opposite direction in the east of the Channel (Stride, 1963). A difficulty arises, however: the sands are not of the same character throughout the supposed drift paths. To the north and south of Brittany and in the Gulf of St Malo stable calcareous sand banks occur which are hydraulic dunes (Hinschberger, 1969; Hommeril, 1967). They are sand formations a few kilometres in length, about 30 m high above bed rock and creased by ripples. They lie in places of decreasing tidal currents, for instance at the mouths of channels (the same phenomenon may occur in shallow off-shore areas) (Guilcher, 1958).

Ripples and dunes are less frequent in the Bay of Biscay, where tidal currents only act on the shallower part of the shelf. The presence of a complex of small sand ridges located at a depth of about 140 m has however been pointed out, south-west of Penmarch (Kenyon and Stride, 1968).

PELITIC SEDIMENTATION

In the English Channel, pelites are found only in bays or estuaries, which are beyond the scope of the present paper. Further south fine sediments are to be found in some areas of the Bay of Biscay (Grande Vasière and fore-coastal muddy areas) and on the continental slope.

The clay minerals of the pelites are illite, kaolinite, chlorite and montmorillonite. Mineralogical and geochemical analyses have shown that the origin of the clay minerals is to be sought in a very old continental formation, first deposited on the shelf and afterwards continually subjected to alteration by the action of the sea, to the point of taking on its present day characteristics (Rumeau and Vanney, 1968, 1969). Therefore, the pelites may be considered as a present day marine deposit, in equilibrium with present day hydrodynamic conditions.

ORGANOGENIC CALCAREOUS SEDIMENTATION

Organogenic sedimentation is especially important in the middle of the Western Channel, in the Celtic Sea and in the Bay of Biscay on the outer part of the shelf, north of 46°N. Algae (Lithothamnion) produce a calcareous gravel in shallow waters, but animals produce the greatest amount of limestone (mollusca, bryozoa, ophiuroidea, balanoglossida, etc.).

Bryozoa abound at the entrance to the Channel where the sea floor is hard (pebbles or rock) and swept by strong currents. Finding their remains in abundance is proof of a drift from a sedimentary productive zone (hard bottom) to sandy zones of accumulation. Reciprocally, the proportion of bryozoic fragments imbedded in the sediment decreases away from the hard bottoms (Boillot, 1965).

Lamellibranchia, on the other hand, live in unconsolidated substrata which they enrich, eventually disturbing its granulometric characteristics. Their present day sedimentary action is very important in the Channel and in the Celtic Sea.

Table 3

	m	Proposed Correlations
Organogenic sedimentation. Pelite deposits in muddy areas. Coastal erosion at 17 m, 40 m, 50-55 m and 100 m below sea-level. Deposit of pebble lines at similar depths. Deposition of the fine sands of the Bay of Biscay. Deposition of organogenic remains contained in the 'reddish-brown sands' and benthic foraminifera actually lying below -120 m (Bay of Biscay).	0 ↑ -120	Flandrian transgression
Settlement of loess and of the second generation of pebbles (English Channel). Cementation of shelly limestone Settlement of the siliceous part of the 'reddish-brown sands' (Bay of Biscay).	-100 or -120	Pre-Flandrian regression (Würm glaciation)
Deposition of the shells of the conglomerate of the Channel and the Bay of Biscay.	0	Pleistocene transgressions
Hollowing out of the karstic depressions in the Channel. Settlement of the first generation of pebbles, in-filling of valleys	-200	Old regressions corresponding to ancient glaciations.
Deposition of calcareous sand on the shelf-edge.	?(+)	Pliocene.
Down-cutting of valleys?	?(-)	Pontian?

Rather than attempt to summarise the studies on present day microfauna, readers are referred to the work of specialists (Le Calvez, 1958; Dupeuble, 1963; Le Calvez and Boillot, 1967; Andreieff and others, 1970).

Recapitulation of Quaternary and Recent Events on the Continental Shelf between the Straits of Dover and the Gouf de Cap Breton

Table 3 clearly shows that the establishment of present day sedimentary conditions is a recent event. Each time a survey of the unconsolidated sedimentary cover has been accompanied by core-sampling, the present day sediments, whether of mud or calcareous sand, have been shown to form a thin veneer covering rocky substratum or coarse siliceous deposits derived from the weathering of continental formations. In general, the unconsolidated cover is thin over the entire shelf. It is obvious that we are witnessing the beginning of a sedimentary cycle.

THE CONTINENTAL SHELF OFF THE SPANISH COAST

South of the Bay of Biscay, the shelf forms a narrow fringe 20 to 30 km wide. Its edge, some 180 m deep, nears the coast at the level of a few major canyons. While the French shelf is, on the whole, flat, the Spanish shelf presents an irregular topography, linked with a very complex structure.

Between Cape Penas and the Estaca Point, rock outcrops are numerous and the sedimentary cover is thin and discontinuous. Along the coast there is a strip of coarse sand which is almost always very shelly. Fine sands, varyingly rich in lutite, cover the rest of the shelf; towards the outer edge they are mixed with dark green grains of glauconite (Gomez de Llarena, 1952, 1955; Lamboy, 1967, 1968), a similar distribution exists towards the east, to the region of Llanes.

From the Estaca Point to Cape Finisterre, the sedimentary veneer is continuous, except on the promontory which extends west, at the north-west of Cape Ortegal and perhaps near Cape Finisterre (Stride and others, 1969). Though more important than in the area situated further east, the thickness of the present day or recent sediments remains rather slight.

The sedimentary areas between Cape Penas

and the Estaca Point continue to Cape Finisterre. But, in the outer part of the shelf, coarse siliceous sand and sometimes gravel, is mixed with fine glauconitic sands. We have therefore to admit that the Spanish shelf, just as the French shelf, retains traces of old continental deposits, perhaps comparable to the reddish brown sands.

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Geology of the East Atlantic continental margin from Finisterre to Casablanca

J. H. Monteiro

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SUMMARY

The published work on this region is reviewed. The following aspects are considered: morphology, sediments, structure and relations with the continent. It is concluded that few investigations have been made and there is a relative lack of knowledge of this region compared with the West Atlantic continental margin and several regions of the East Atlantic continental margin, mainly the North Sea and the Bay of Biscay.

SOMMAIRE

Les travaux publiés sur cette région sont examinés. Les aspects suivants sont considérés: la morphologie, les sédiments, la structure et les rapports avec le continent. On estime que peu d'études ont été faits, et on manque de connaissance sur cette région en comparaison avec la marge continentale de l'Atlantique de l'ouest, et avec plusieurs régions de la marge continentale de l'Atlantique de l'Est, surtout la Mer du Nord et le Golfe de Gascogne.

ZUSAMMENFASSUNG

Man bespricht die herausgegebenen Werke über dieses Gebiet. Man betrachtet die folgenden Seiten des Themas: Morphologie, Sedimente, tektonische Struktur und Beziehungen zum Kontinent. Man entscheidet, dass wenige Untersuchungen stattgefunden haben. Es gibt einen vergleichsweise Mangel an Kenntnisse von diesem Gebiet im Vergleich mit dem Westatlantischen Kontinentalrand und mit mehreren Gebieten vom Ostatlantischen Kontinentalrand und mit mehreren Gebieten vom Ostatlantischen Kontinentalrand, hauptsächlich mit der Nordsee und dem Meerbusen von Biskaya.

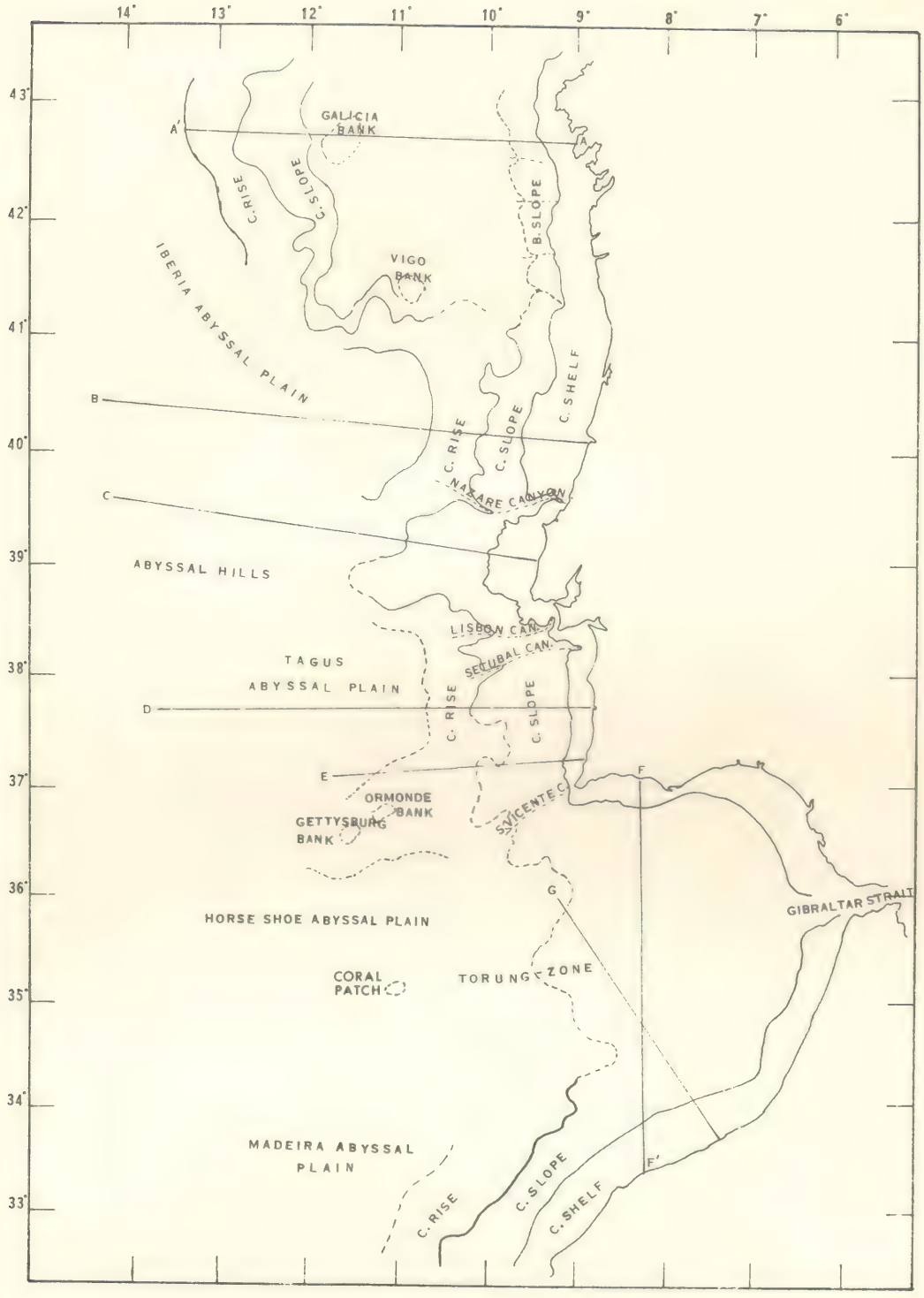

Fig. 1. Principal morphological regions (based on Laughton's charts, N.I.O., 1967).

Geology of the East Atlantic continental margin from Finisterre to Casablanca

J. H. MONTEIRO

INTRODUCTION

This paper reviews the geology of the continental margin off the Iberian Peninsula and North-west Africa, between Cape Finisterre and Casablanca. Most of the review was based on published papers, but compilations of geological observations were kindly given by Dr. Roberts of the National Institute of Oceanography and Dr. Seibold of the Geologisch und Paläontologisches Institut der Universität Kiel. A compilation of all available observations on 1/250 000 GEBCO Charts was made at the Portuguese Hydrographic Institute, using one chart for geophysical data and another for geological observations.

Most of the published papers on the subject are fairly recent. The bibliography contains fifty-nine papers referring directly to the region; six were published before 1940; none between 1940-49; fourteen between 1950-59; thirty seven between 1960-69; and two in 1970.

Our present revision is based on that compilation and will be divided into the following parts: morphology; sedimentology; rocks older than Holocene; structure; relations with the continent; conclusions and suggestions.

MORPHOLOGY

A general description of the entire continental margin from the coast to the abyssal plains can be made using the GEBCO 1/1000 000 charts 042, 043, 060, 081, with bathymetric contours compiled from P. D. R. records by Laughton of the National Institute of Oceanography. The relief of the continental shelf and the upper slope off the Iberian Peninsula are shown on charts compiled by Berthois and Brenot on approximately 1/230 000 scale. These charts are based on 1952 to 1964 surveys by the French vessels 'President-Theodore-Tissier' and 'Thalassa', plus a compilation of the Portuguese Hydrographic charts. The charts, Nos. 1 to 6, published by the Portuguese Hydrographic Institute on approximately 1/150 000 scale, also have discrete soundings on the entire shelf. For the continental margin off North-west Africa detailed studies exist by Bremonde (1953), Lacombe (1955), Grousson (1957), Gougenheim (1959).

Several authors have been interested in the morphological aspects of the continental margin; Ramalho (1932) described the principal features of the relief shown by the Portuguese

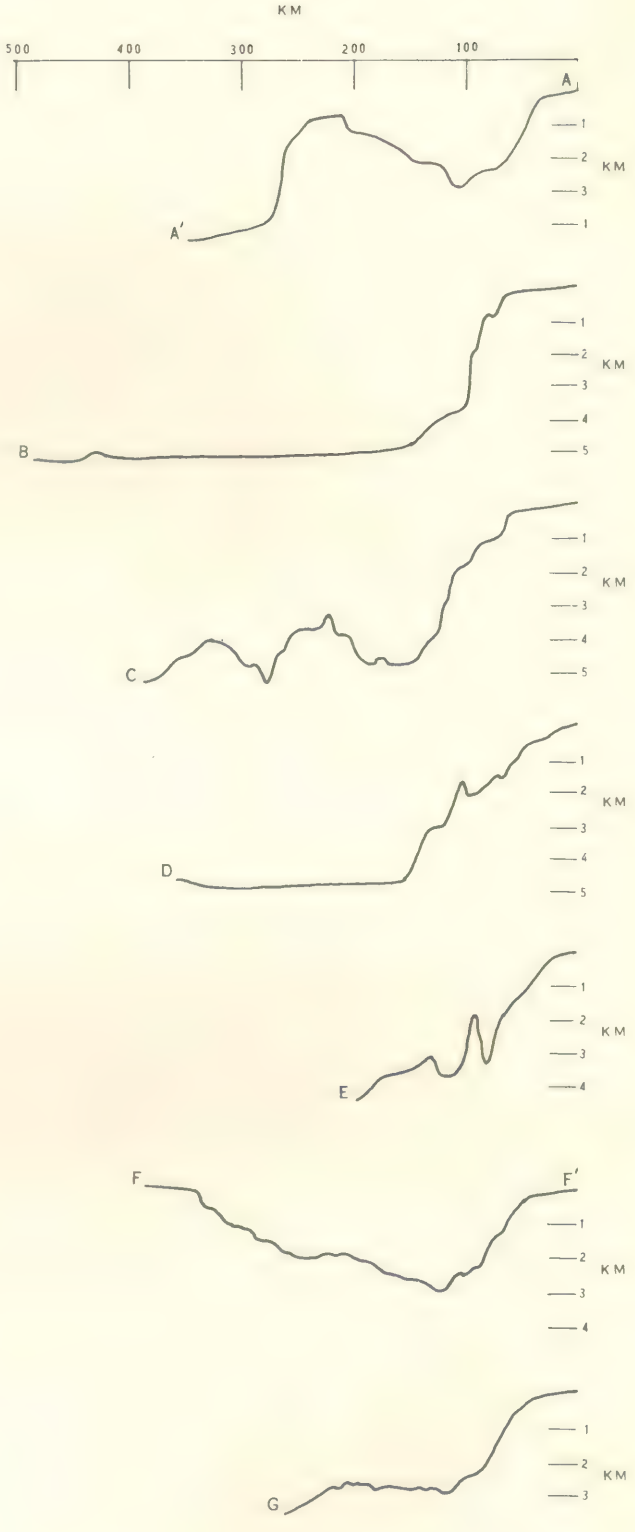

Fig. 2. Profiles of the continental margin (from Laughton's charts, N.I.O., 1967).

TABLE 1

PROFILE	Continental Shelf			Continental Slope			Continental rise			Total width of the margin (km)
	width (km)	depth of the edge (m)	slope °/oo	width (km)	depth of the base (km)	slope °/oo	width (km)	depth of base (km)	slope °/oo	
A		-								
42° 40'	30	180	6	38	2.2	55				
A' GALICIA BANK				34	3.8	105	43	4.4	1.3	31.6
B 40° 05	59	180	3	41	3.6	83	47	4.7	2.4	147
C 39° 02	61	290	5	63	3.7	58				144
D 37° 40'	24	255 ?	11	134	4.7	33				150
E 37° 15'	28	290 ?	10	50	3.5	64				78
F ALGARVE	48	181	4	54	1.7	28				102
F' MARROCOS	46	218	5	74	3.0	38				120
G 33° 40'	41	145	4	68	2.9	40				109
CURRAY (1968)	75	130	1.7	20-100	1.5-3.5	70	0-600	4	1-10	

Explanation of Table 1

From Finisterre to 41° N (Profile A). Shelf: narrow shelf only 20 to 40 km wide; the edge of the shelf varies in depth, from 200 to 400 m. Slope: there is a basin slope, with 55 °/oo gradient, between the coast and the Galicia Bank and a steeper continental slope with 105 °/oo gradient, from Galicia Bank to the Iberian abyssal plain. Canyons: canyons cut on the shelf edge are not related to the present drainage of the continent. Rise: exists seawards of the Galicia Bank. Other features: the existence of the offshore banks allows this region to be considered as a borderland.

From 41° N to the north wall of the Nazaré Canyon (Profile B). Shelf: closer to average width (75 km), attaining a width of 62 km. The edge of the shelf is about 200 m deep. There is a depression in the shelf off Aveiro. Slope: well-defined slope, 40 km wide, base at 3.6 km and dipping seaward with a 83 °/oo gradient. Rise: well-defined, 47 km wide, with the base at 4.5 km depth and dipping seaward with a 2.4 °/oo gradient. Canyons: not present. Other features: Berthois and others (1965) suggest the occurrence of submarine deltas in front of the Douro and Mondego and a change in the strike of the shelf at 40° 32' N.

From the north wall of the Nazaré Canyon to the south wall of the Setubal Canyon (Profile C). Shelf: about 60 km wide with a marked relief of the 100 m contour which prolongs the Sintra batholith. South of Cabo Raso the shelf is narrower. Slope: difficult to delimit due to the continuation of a zone with oceanic relief, and with several benches. Canyons: three important canyons: the Nazare Canyon cuts very close to shore; it is 170 km long and up to 4.5 km deep, it lies near a land point. The Lisbon canyon has two heads cut on the shelf edge south of the Tagus Estuary; both appear to be migrating from south to north probably adjusting to the present bottom currents. This agrees with Belderson and Stride (1969) who, from Asdic records, described active cutting of canyons in several places, along the East Atlantic continental margins. The Setubal Canyon heads towards a straight beach south of Cabo Espichel. The Berthois and others charts show the Lisbon and Setubal canyons to join at a depth of about 1.8 km, but on the Laughton charts they appear separated, even at 4.5 km; this reveals the necessity for more detailed sounding of the region. The Nazaré Canyon contributes sediments to the Iberian abyssal plain and the Lisbon and Setubal canyons contributes sediments to the Tagus abyssal plain.

From the south wall of the Setubal Canyon to Cabo de S. Vicente (Profiles D, E). Shelf: narrow, 15 to 20 km wide, steep slopes 8 °/oo to 11 °/oo. Slope: width 140 to 150 km, sloping 34 °/oo. Canyons: S. Vicente Canyon. This canyon extends in a north-east to south-west direction west of Cabo de S. Vicente. Soundings of the canyon do not extend beyond 2.1 km. Other features: the southern part of this region is complicated by the occurrence of the eastward continuation of the Gettysburg and Ormonde banks and the Azores Gibraltar Ridge.

Southern part of the Iberian Peninsula and the region in front of Gibraltar (Profile F). Shelf: narrow shelf, 48 km wide, off Portugal widening near Cadiz to 70 km. There is no continental shelf in the region of the strait. Slope: wide continental slope, steep slopes are only locally developed. Several benches are present. Canyons: a series of small submarine canyons which cut both the slope and the shelf, and probably are related to the rivers Guadaleti, Guadalquivir, Tinto Odiel, Guadiana and Odelouca. Other features: existence of slump structures was mentioned by Roberts and Stride (1968). The minimum sill depth of the Gibraltar Strait is 268 m.

Gibraltar to Casablanca (Profile G). Shelf: the continental shelf is 41 km wide in the profile G, the shelf edge is about 150 m deep. Slope: well-developed continental slope, base of the slope at 2.9 km, width 68 km and dipping seaward with a gradient of 40 °/oo. Rise: present south of Casablanca.

hydrographic charts; Andrade (1937), using the same charts, described the submarine canyons off the Portuguese coast; Giesel and Seibold (1968) gave a general morphological description of the southern Portuguese Continental margin and the Gibraltar sector; Johnson (1965) reviewed the morphological features of the Gibraltar region and the relief of Gibraltar Strait was studied by Hernandez Pacheco (1961). McMaster and Lachance (1969), show representative sounding profiles of the continental shelf and upper slope off North-west Africa. Heezen and others, (1959, pp. 37-38) reviewed the Iberian, Gibraltar, and north-west Africa continental margins.

A diagrammatic profile of the continental terrace shown by Curray (1966, p. 208) was compared with the different profiles from the Iberian margin, taken from the Laughton chart. Several units can be recognised (Table 1, Figs. 1 and 2).

SEDIMENTOLOGY

Iberian Sector

The first reference to the sediments comes from the 'Challenger Reports', Murray and Renard (1891, p. 148) reporting on the voyage from England to Gibraltar. The existence of green mud and green sands near the Portuguese coast was mentioned.

In the Galicia region 160 sediment samples were collected during the 'Xauen' expeditions of

1949-1950 (Gomez de Llarena 1952). Inside the 'rias' and from shore to 50 m depth he reported black mud with a very strong odour of hydrogen sulphide, implying anoxic conditions and on the outer shelf sands and shelly sands. Some of the sands have foraminiferal moulds of glauconite. Perez Mateos (1952), considers the heavy mineral assemblages present in the samples to be derived from the acidic rocks of the coast of Galicia. Colom (1952), reported the study of the recent foraminifera of the same samples. Lamboy (1967) reported the results of the study of 160 dredge samples from the north-west of the Iberian Peninsula, and the occurrence of two types of glauconite of bathymetric significance. He reported values of 65 per cent of glauconite in the fraction $> 50\mu$ of the samples, for the region between 160 and 200 m depth. Lamboy (1968) explains the formation of glauconite by epigenesis from the calcareous debris.

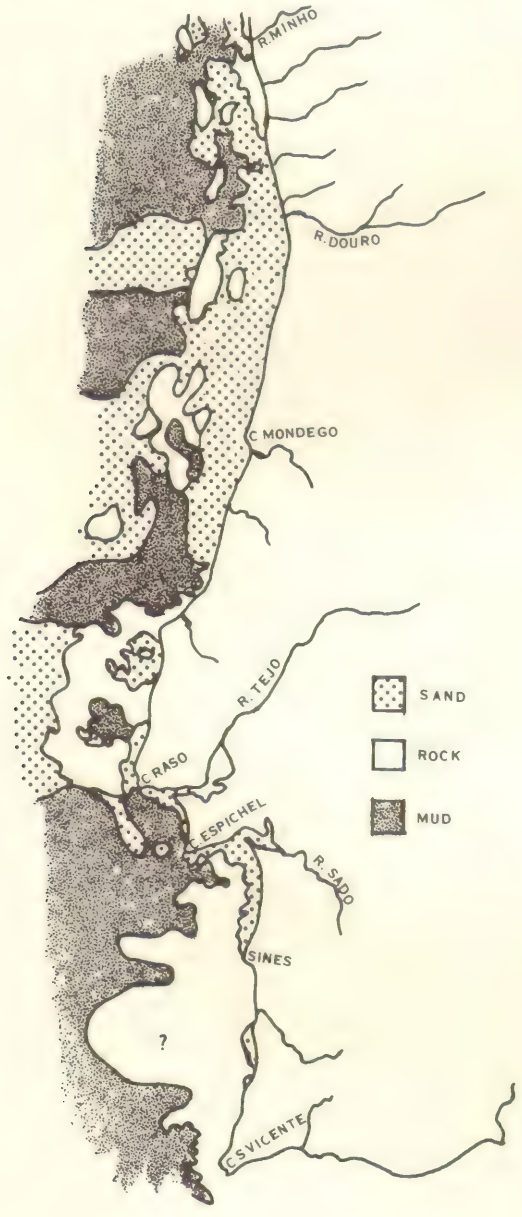

Fig. 3. Sediments of the continental shelf off Portugal (from Cartas Litologicas Submarinas, Instituto Hidrográfico, Lisbon).

Few sediments samples were collected during the expeditions of the 'Xauen' in 1955-56. Gonzalez Paña (1958) and Pérez Mateos (1958) respectively studied the clay minerals and the heavy minerals of the samples. They found, for the Galicia shelf, a high proportion of illite, kaolinite and halloysite, and the sands are interpreted as being derived from a source in the Galicia coast.

Sediment charts, 'Cartas litológicas submarinas', were published between 1915-1930, for the continental shelf off Portugal. Gomes (1919) and Nobre (1929) described the sediment samples and Gomes (1915-1916) the methods used.

From the Minho River to Espinho, litoral sands extend from the coast to 90 m depth. Patches of sand and gravel also occur near 200 m depth; shell fragments are relatively scarce in this region. The charts also show elongated patches of mud mainly between the Lima and Douro rivers. Gomes (1915-16) reported quartzose sands and biogenic sediments. From the Douro to the Mondego River the sand extends to a depth of more than 100 m and there is also a zone of sand around the 200 m contour. Outside this contour there is a zone of mud.

South of Cabo Mondego the zone of sand is narrower following the 100 m contour, and rocky bottom occurs. South of the Nazaré Canyon, sand is scarce in the littoral zone, but there are abundant patches of gravel and biogenic sediments. Indications of rocky bottom are also frequent in this region.

Between Cabo Raso and Cabo de Sines there is a littoral zone of sand from shore to 50 or 100 m depth. Between 50 to 150 m depth there are patches of mud and sandy mud. Seaward of the 200 m contour the bottom is mainly rocky. South of Cabo Espichel there are abundant patches of shells and gravel. In a few places around 200 m depth there are muddy sand patches. Inside the canyons there is mud with patches of sand.

South of Cabo de Sines there is a zone of littoral sand from shore to 50 m depth and an outer zone of sand between 200 m and 300 m depth, and mud outside this zone. There are frequent signs of rocks.

On the continental shelf off the southern coast of Portugal there are, from Cabo de S. Vicente to Cabo de Sta Maria, a narrow littoral zone of sand, frequent evidence of rocks, and abundant patches of shells. East of Cabo de Sta Maria there is an increase of sandy mud extending to the 500 m contour. There is a zone of littoral sands and there is also a zone of biogenic sands

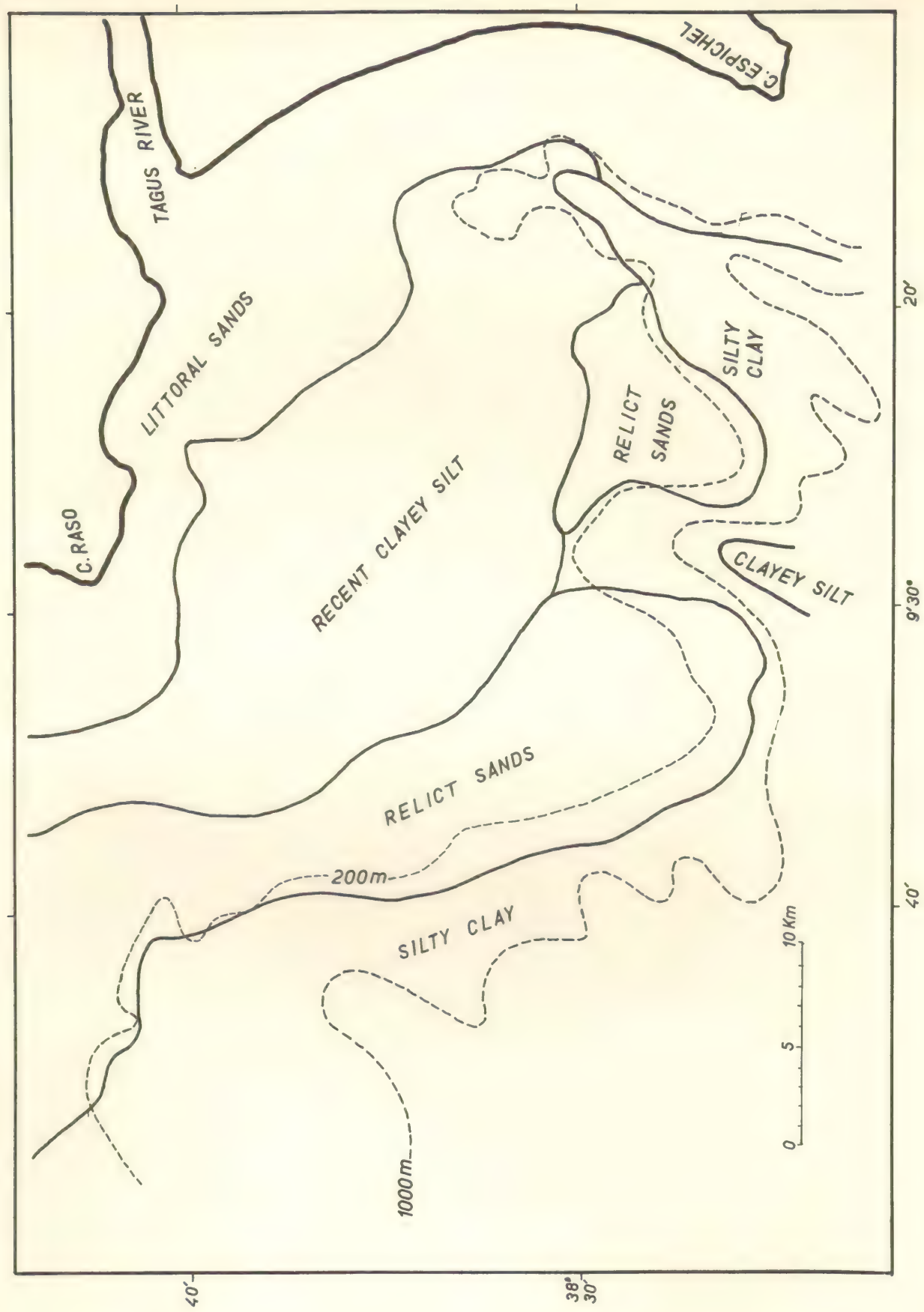

Fig. 4. Sediments of the Portuguese continental margin off Setubal Peninsula.

close to the 200 m depth contour.

Andrade (1937, p. 142) compiled lithologic charts showing the distribution of sand, mud and rocks. Recently the Portuguese Geological Survey published a 1/1000 000 chart of the Quaternary rocks which also contains a compilation of the 1930 charts.

The continental shelf off the Tagus River mouth was recently studied in detail at the Portuguese Hydrographic Institute, and four main sedimentary facies were outlined (Fig. 4): littoral terrigenous sands near the coast; basal transgressive sands, near the edge of the shelf; recent clayey silts carried in suspension by the Tagus River, separating the two sand facies; and an upper slope facies of recent silty clays. The outer sands are rich in calcium carbonate and glauconite. The study of calcium carbonate mineralogy revealed high percentage of low-magnesian calcite, aragonite and some dolomite.

The 'Meteor' Cruise 8/67 also collected sediment samples. Kudrass and Thiede (in press) have already reported the sand grains analysis and the study of planktonic foraminifera from two cores of the continental slope off Portugal and Morocco. They found a good correlation between the percentage of terrigenous material, pteropod content, grain size of foraminifera and coiling ratios of Globorotalia. They also reported sediments from the last Ice Age in the lower parts of both cores.

The R.R.S. 'John Murray' (Applied Geochemistry Research Group, Imperial College) collected bottom samples in three east-west traverses between Cabo de Sines and Cabo de S. Vicente.

Nutter (1969) reported the study of the sediments from the three profiles. The sediments are mainly sands. From the shore to the ocean he found littoral coarse sand; fine sand; a belt of coarse sand about 200 m depth; and fine sand and sandy mud on the upper continental slope. Calcium carbonate distribution in the area shows high percentage of calcium carbonate (> 75 per cent) in the south and a low one in the north. The littoral sands are quartz sands. The outer shelf sands contain a high proportion of shelly material and some glauconite; they appear to be relict sands.

Zone in front of Gibraltar

Johnson (1965, p. 15) compiled data from various sources for this region; his chart shows a zone of littoral sands, narrow off the Portuguese Coast, that widens between Huelva and Cadiz.

There is abundant rocky bottom in the west part of the Portuguese Coast, in the coast south of Cadiz and in the Gibraltar Strait. In front of the strait a tongue of sand and gravel extends west-north-west. Mud patches extend from the mouths of the Guadiana and Guadalquivir rivers to the west.

Gonzalez Peña (1958) and Pérez Mateos (1958) studied a few samples from the Andalusian coast and reported kaolinite, illite, and halloysite for the clay minerals. The sediments are rich in fines because of the proximity of Guadalquivir. They concluded that the reported sands originated from the adjacent continent.

Northwestern African continental shelf from Gibraltar to Casablanca

A total of 160 grab samples were collected by the R/V 'Trident' (McMaster and Lachance, 1969). The shelf is covered by sand with silt admixtures.

The highest percentage of sand occurs in the outer shelf. The sands are fine sands composed of shell fragments, faecal pellets, foraminifera, glauconite and quartz. Glauconite occurs in several samples of the shelf and upper slope reaching 5-95 per cent of the sand fraction. Mid-shelf belts of silt lie between the coast line bulges; the northernmost belt contains 25-49 per cent clay which is composed of illite, kaolinite and some degraded illite and chlorite. The Sebou River brought the fines to the midshelf. North-west of Morocco, the currents near the strait prevent the occurrence of a belt of fines on the mid-shelf. Bell and Goodell (1967) suggest a detrital origin for some of the glauconite. Zones of low terrigenous supply are rich in biogenic deposits, reaching 50 per cent of $CaCO_3$.

Tooms (1968) reported a traverse off Cape Fedala in which he sampled mainly muddy sediments.

The report on cruise 69/1 of R.R.S. 'John Murray' by Tooms (1969) mentions two traverses (18 and 19) north of Casablanca with dredging and gravity coring: from shore to 53 m depth sand and fine sand were reported, beyond 53 m depth a zone of muddy sand; around 500 m sand and sandy mud were found. Analysis of sediments samples made aboard the ship disclosed a zone of abnormally high phosphate content at a depth of approximately 200 m on the Moroccan shelf. High phosphate contents were also observed in the vicinity of the phosphatic rocks and at a number of localities close inshore.

Sediment transport

Longshore drift: the predominant direction of wave attack is approximately west-north-west (Reis de Carvalho and Barcelo, 1966), which results in a predominant drift from north to south on the west coast of Portugal, and west to east on the coast of Algarve.

Studies with 'labelled sand', sedimentological beach studies and model studies (Abecasis and others, 1962) show selective transport from north to south. A historical study of the port of

Aveiro (Abecasis, 1955), gave historical data from 1318 through to 1945 on the closing of the lagoon entrance. In the tenth century a sandy spit started to proceed southwards from Espinho, revealing the north-south movement.

The bottom currents on the west coast of Portugal and North-west Africa are unknown. The surface currents are from north to south as part of the Canaries current (Wooster and Reid, 1963), but there are possibilities of a north-flowing counter current (Rodrigues, D., personal communication, 1969). Those currents and unknown gyres may be responsible for the transport of fines in suspension. Giesel and Seibold (1968) suggested the probable importance of variations of the water movement, and of the interaction of water movement and relief, for filling depressions observed around obstacles in the southern part of the Portuguese continental shelf. They also suggested that the small thickness of sediment from Cabo de S. Vicente to approximately 37° 20' N was due to the action of the flow of the Mediterranean water. The sediments in the strait will be eroded and subject to continuous transport. The current turns to the north-west due to the Coriolis force and will act on the southern slope of the Portuguese continental margin.

A study of submarine photographs of the bottom of the Gulf of Cadiz by Melieres and others (1970) suggests the importance of the currents in the morphology of this region and their relation to the system of currents of the Gibraltar Strait.

Evidence for the movement of sediments from the continental margin to the deep-sea abyssal plain has been given by Keen (1963) for the Iberian Plain and by Duplaix and others (1965) for the Tagus abyssal plain.

During the dives of the F.N.R.S. in the Setubal canyon Pérès and others (1957) reported down-canyon currents and observed rock surfaces which revealed erosion or at least no deposition.

The actual availability of sediments from terrigenous sources for the margin, is relatively low. The principal rivers are the Douro, Tejo, Guadiana and Guadalquivir. Griffin and others (1968) found the area adjacent to the Iberian Peninsula to have very high concentrations of illite (70 per cent in the $< 2\mu$ fraction). They suggested that the sediments from the small rivers, draining the Iberian Peninsula, have very minor influence compared to the aeolian components.

ROCKS OLDER THAN HOLOCENE REPORTED FROM THE CONTINENTAL MARGIN BETWEEN FINISTERRE AND CASABLANCA

Wiseman and Ovey (1950, p. 42) found reworked Upper Cretaceous foraminifera in sediment from a depth of 2743 m off Cape Finisterre: Funnel (1964) reported Upper Maestrichtian in chalk from Galicia Bank. Black and others (1964) reported pellet limestones exposed in the flanks of Vigo Seamount which resemble pellet limestones of Jurassic age outcropping in Portugal.

Sutton and others (1957) reported Miocene sediments from a depth of 1280 m in the wall of the Lisbon Canyon. A dredge sample collected by the Portuguese Hydrographic Institute on the continental slope west of Cabo Espichel also seems to be Miocene in age (Ésteves Matos, personal communication, 1968).

Asdic (side-scan sonar) lines were employed to search rock outcrops and strike lines on R.R.S. 'Discovery II' and R.R.S. 'Discovery' between 1958 and 1967. Stride and others (1969) presented an acoustic map of the continental shelf located 25 km oceanward of the Jurassic rocks of western Portugal; it showed three large outcrops of sedimentary strata. An acoustic map of the sea floor on the continental shelf off north-west Spain shows massive rock which is probably igneous.

Dredge samples reported by Tooms (1969) are composed of limestone, mudstone, sandstone and slate but the detailed study is not yet available.

The occurrence of Palaeozoic crystalline rocks is observed in Berlengas and Farilhões islands (Andrade, 1937) near the south wall of the Nazaré Canyon.

STRUCTURE

Seismic reflection

Before 1969 very few profiles had been made but in that year several oil companies made a general survey of the Portuguese continental shelf; their results have not yet been disclosed.

Stride and others (1969) studied a reflection Arcer profile off Finisterre (profile 15). The eastern end of the profile revealed a rough surface overlain by westerly dipping strata of Mesozoic or perhaps Tertiary aspect, upon which there is a thin development of horizontal material. Westwards the profile includes the Galicia Bank. Late Cretaceous limestones capping the top of Galicia Bank have a crenulated bedding. Their style of folding may be the same as that in Portugal and their softness is in accordance with their low apparent velocity. The almost uncontorted bedding and considerable lateral extent of the upper sediments on both sides of the bank, suggest that they are probably Tertiary strata with a thin cover of younger material. There are abundant indications of faulting.

Other profiles studied by Curray and others (1966) and Stride and others (1969) are an Arcer (profile 16) and an Air Gun (profile 12), approxi-

mately off Lisbon. Profile 16 shows that well-bedded rocks of Tertiary aspect extend from the outer part of the continental shelf down the upper part of the continental slope and are underlain by undifferentiated rocks. A prominent bench on the slope was interpreted as an outcrop of basement type rock. The deep water portion of the profile shows a thick section of well bedded material which may be Quaternary, underlain by an interface with signs of erosion. This interface may indicate the top of Tertiary material. The Air Gun profile revealed that the continental slope was a good reflector but failed to detect any bedding beneath it. This suggests that basement type rocks extend to a depth of at least 2.7 km.

Giesel and Seibold (1968) studied Air Gun profiles from the continental margin of southern Portugal and North-west Africa. The reflection profiles of southern Portugal showed that the unconsolidated sediments fill depressions in the basement. Due to the relief of the basement there are patches of sedimentary beds of different thicknesses. The basement outcrops 50 km south-west of Cabo de Sines and there are other basement outcrops of less importance. The Lisbon and Setubal Canyons cut the basement and there are sediment fills in the Setubal Canyon. The basement may be of Tertiary, Mesozoic or older rocks. On the Moroccan shelf, basement outcrops frequently. The internal stratification of the basement dips 1^{o} seaward but the beds are discordant with the shelf and slope.

On the shelf the discordance may be due to erosion by waves and currents but in the slope may be the result of tectonic activity.

In the west of Gibraltar Giesel and Seibold (1968) found no clear indications of the basement-sediment interface. The same authors interpreted the series of highs and lows 'Rinnen' and 'Rucken' as erosional depressions more or less filled by sediments, but interpretation as salt 'stocks' is not definitely excluded. Roberts and Stride (1968) studied seismic reflection profiles of a series of long sub-parallel ridges with a westerly to west-south-westerly trend on the continental slope south of Portugal. The profiles show folds closely related to the relief. The authors interpreted the folds as resulting from down slope movement and crumpling against the submerged edge of the Betic block, during the Pliocene uplift of Iberia. Nutter (1969) reported three reflection profiles between Sines and S. Vicente made during the 1969 cruise of the R.R.S. John Murray. The profiles showed thin sediment cover. A marked unconformity was present between folded older rocks and the more recent sediments.

McMaster and Lachance (1968) analysed several profiles off Morocco. The seaward dipping reflectors show progradation of the continental shelf but contorted layers and normal faulting reveal that tectonics have had a major effect on the structure. Erosional surfaces and post-erosional deposition are also recognised.

Seismic refraction

Seismic refraction results reported from Galicia Bank (Black and others, 1964) revealed 0.4 km of uncemented limestones overlying 4.2 km of hard limestones that rests on a crystalline basement.

Only seismic refraction investigations have been carried out in the Gulf of Cadiz (Drake and others, 1959). Drake and others found four horizons of 1.85 km/sec; 2.37 km/sec; 3.67 km/sec; 5.64 km/sec which were correlated with the formations found on the continent; respectively, Quaternary, Pliocene, Miocene, Mesozoic calcareous rocks; Keuper; Palaeozoic basement.

Gravimetry and magnetism

Loncarevic (in Black and others, 1964) did sea gravimeter observations over Galicia Bank in 1960 and observed no substantial mass defect. The magnetic survey over the three seamounts showed they were non-magnetic.

Dr. Roberts (personal communication, 1970) mentioned that the N.I.O. is processing gravimetric data from the zone in front of Gibraltar. Gravity anomalies were found in continuations of the arcs of southern Spain and northern Morocco.

Allan (1965) reported a high magnetic anomaly in the total-magnetic intensity off the mouth of the River Tagus. Recent work by the Hydrographic Institute and the Geophysical Institute of the University of Lisbon, off Sintra and Sines, confirmed the large anomaly and revealed anomalies near the edge of the shelf.

RELATIONS WITH THE CONTINENT

The principal structural units in the adjacent continental area of the region under description are: the Hesperian massif covering north-western Spain and extending into southern Portugal as far as Algarve; the Mesozoic belts of west and south Portugal; the Betic Cordillera in southern Spain; the Riff and the Moroccan Meseta in north-west Africa.

The structural grain of the Hesperian massif appears to be unrelated to the adjacent continental margin. Stride and others (1969) associate this with the rupture and drift of Europe from North America and the anti-clockwise rotation of the Iberian Peninsula.

The relations of the structural units of the continent with the continental margin are by no means clear. The $^{1}/1\ 000\ 000$ tectonic chart made by the Portuguese geological survey shows the

principal known tectonic structures in the continent: their sub-sea prolongations are unknown but a great number of them may continue to the continental margin.

Giesel and Seibold (1968) found no continuation of the structures of the Guadalquivir zone. They suggest that the zone with 'nappes de glissement' with 'olistostromes' coming from the south-east in southern Spain may be recognised in the profile XXVII. The Betic zone and the tectonic units of Gibraltar are very complex and it is impossible to investigate their submarine continuations with the methods used. They suggest a series of seismic profiles as a promising method to resolve the problem of whether the arcs of southern Spain and northern Morocco constitute a single unity or not.

The Gibraltar strait appears to be a zone of subsidence (Giermann, 1961; 1962); the tectonics are complex and the extrapolations inferred by Hernandez Pacheco (1961) from the land geology are not yet confirmed. Glangeaud (1962) suggested that the strait may be related with a zone of 'cassures fondamentales' coming from Azores through Gibraltar and going to the Tyrrhenian Sea. Krause (1965) considers a lineation between Azores and Gibraltar (East Azores fracture zone) as being the displacement along this left-lateral fracture.

A compilation of earthquake epicentres made at the Geophysical Institute of the University of Lisbon shows accumulation of epicentres E and NE of Gibraltar and SW of Cape of S. Vicente, but not in the direct continuation to WSW of the Gibraltar Strait.

Le Pichon (1968) applying the concepts of sea floor spreading and plate tectonics, indicates compression between the European plate and the African plate, at a rate of 1.5 to 1.9 cm/year along the Azores Gibraltar ridge.

The formation of the Bay of Biscay as result of the relative anti-clockwise rotation of Iberia (Van der Voo,1969; Watkins and Richardson, 1968; Jones and Ewing,1969) or by the relative westward translation of the Iberian Peninsula (Mattauer, 1969a and b) may have different implications in the adjacent Continental margin.

CONCLUSIONS AND SUGGESTIONS

Morphology

1. There is a region north of 41° N with the characteristics of a borderland.
2. Near 37° N and 39° N it is difficult to define the seaward limit of the continental terrace, due to the existence of zones of oceanic relief.
3. In the Gibraltar sector steep slopes are only locally developed in the continental slope. Several benches are present.
4. The continental shelf in general is a narrow shelf with less than the average width (75 km) (Curray, 1968). The shelf edge is of varying depth indicating tectonic instability.
5. The four large canyons off the Portuguese coast are practically unexplored.
6. The published work on the morphology of the Iberian Continental margin is based on a large scale mapping and old methods. There is a need for more detailed morphology of the Iberian continental margin.

Sediments

1. The terrigenous sediment supply for the continental margin from Finisterre to Casablanca is relatively low.
2. There is an abundance of glauconite, revealing zones of low sedimentation, probably associated with relict sediments from the Holocene transgression.
3. Zones of high calcium carbonate are present locally, probably associated with zones of high productivity or low terrigenous supply.
4. Turbulence near the edge of the shelf may be associated with the sands present near the edge of the shelf.
5. The system of currents associated with the Mediterranean water is important in erosion, transport, and deposition of sediments in the Gulf of Cadiz and in the southern continental slope of the Iberian Peninsula.
6. A belt of mid-shelf muds is present in regions where rivers supply fines.
7. Relative enrichment in phosphate occurs at least in the Moroccan shelf.
8. The Iberian margin may be tentatively classified as a sedimentary basin with low terrigenous supply, moderate tectonic instability, and it is open to marine reworking.
9. A detailed study of sediments is necessary for the understanding of the Holocene sedimentation in the Iberian shelf and slope.

Rocks older than Holocene

1. Very few samples have been recovered and no general interpretation of the lithology and stratigraphy of the continental margin between Finisterre and Casablanca is yet possible.
2. The oldest rocks sampled were Jurassic pellet limestones from the Vigo seamount.
3. Because of the low rate of sedimentation, and because there is evidence of rocky bottoms in charts, the continental slope of Portugal will be a promising place for dredging rock outcrops. Side-scan sonar surveys may be of great help in locating rock outcrops.

Structure

1. Reflection profiles off the Iberian margin show little accumulation of sediment. The rocks of the continental terrace are folded and faulted.
2. The steep continental slope seems to be of tectonic origin.
3. Profiling in north-west Africa shows progradation of the continental shelf.
4. Interpretation of the profiles made by the oil companies of Portugal will probably define the sediment filling of the continental shelf basement relief and the relations with the continent.
5. Reflection profiling of the slope made in conjunction with rock dredging is a promising method to study the geology of the continental terrace.
6. There have been few investigations of the deep structure in the Iberian continental margin. This is evident in the charts published by Worzel (1968) which demonstrate the almost complete lack of data in the Iberian region, compared with the Atlantic margin of the United States, or the North Sea.

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